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Abstract

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SOUTHEASTERN GEOLOGY

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POSSIBLE RELICT GLACIAL FEATURES IN THE BLACK BALSAM KNOB AND RICHLAND BALSAM AREA, NORTH CAROLINA

By

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ABSTRACT

Several large and very deep north and east facing ravines in the highest mountains of southwestern North Carolina at 35° north latitude could have possibly contained very small alpine ice masses or at least deep snowfields. Small ice masses may have developed in the upper portions of the ravines above 4800 feet (1218 m). Striated quartz veins have been found at two locations on Black Balsam Knob above 5800 feet (1768 m) but these striae could very well be the result of faulting. The upper reaches of the largest ravines that face east, southeast and north have cirque-like forms that have been modified by mass-wasting processes. Abundant bouldery talus and block streams demonstrate a past cooler climate and suggest the treeline was lower.

INTRODUCTION

The Black Balsam and Richland Balsam area is located along the crest of the highest Blue Ridge mountains in Haywood County, North Carolina (Figure 1) and is paralleled for much of its length by the Blue Ridge Parkway. The area is encompassed by the Sam Knob and Shining Rock 7.5-minute quadrangles. The highest peak in this region, Richland Balsam, reaches an elevation of 6410 feet (1839 m) and there are several other crests above 6000 feet (1827 m). Black Balsam Knob, east of Richland Balsam, is part of a north-south ridge and is the highest point on the ridgecrest at 6214 feet (1831 m) above sea level. This region is drained by the east and west forks of the Pigeon River.

Bedrock in the area is a complex of Precambrian gneiss, schist and migmatite. For details of structure and lithology see Hadley and Nelson (1971). Gneiss on Black Balsam Knob is cut by numerous quartz veins. The foliation of the bedrock is complexly folded, but generally

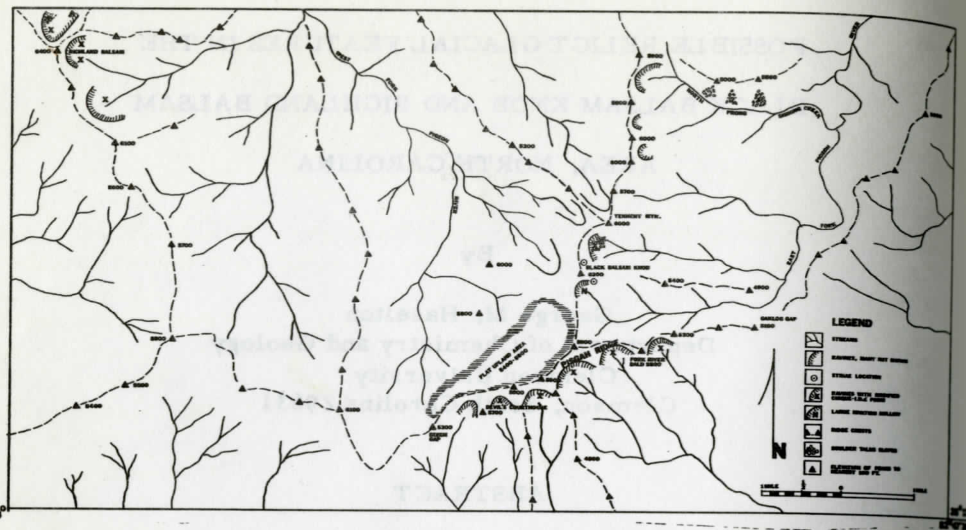

Figure 1. Index map of Black Balsam Knob, Richland Balsam area, North Carolina.

strikes north 50 to 70 degrees east, and dips steeply to the northwest. The northwesterly dip provides the structure for the development of cuesta-shaped hills. The very steep eastern side of the peaks in this area, especially between Devil's Courthouse and Oaklog Gap (Figure 1), is cut by deep ravines. The uppermost reaches of some of these ravines are cirque-like in shape although modified by stream erosion and mass wasting. These are best seen just one mile south of Black Balsam Knob.

Acknowledgments

The writer is greatly indebted to William A. White of the University of North Carolina, G. Michael Clark of the University of Tennessee, Paul K. Birkhead of Clemson University and James O. Berkland at Appalachian State University for their valuable editorial comments and other timely suggestions.

GENERAL SETTING

The main area of the present field study has been carried out in the vicinity of Black Balsam Knob. This is a dome-shaped peak with steepest slopes on the east side. The upper portion of the mountain is a grassy lawn appearing to have a modern day timberline, but the lack of trees above 6000 feet (1827 m) is the result of a severe forest fire that occurred in November, 1925. Small fir trees are beginning to re-establish themselves close to the uppermost parts of the summit.

CLIMATE

Climate at this altitude is subalpine with cool summers and moderately cold winters. Snowfalls are not uncommon from November through April and during some winters a snowpack of 3 to 4 feet 1-1.2 m) occurs at elevations above 5000 feet (1520 m) with drifts above automobile level. Presently, the first freezing temperatures occur during the first week in October on Mount Mitchell (30 minutes latitude north of and 400 feet (121 m) higher than this field area) and the last freezing temperature is June 1 (U. S. Weather Bureau, 1965). Therefore, a modern-day freeze-thaw cycle can occur at the higher elevations for 8 months of the year. How much of the total precipitation falls as snow in the Black Balsam Knob area is not known. Mr. Danny W. Hile, District Ranger (written communication, 1968) indicates that a "ball-park estimate" would be 10 to 20 percent. If this is true, then 75 to 150 inches (1905-3810 mm) of snowfall occur here. It is of interest to learn (U. S. Weather Bureau, 1965) that at Grandfather Mountain, where evidence for alpine glaciation is believed to have been found, (Berkland and Raymond, 1973) the yearly precipitation values, 60 inches (1524 mm), are almost identical to those inferred at Black Balsam Knob. Also of interest are snowfall data from Banner-Elk, North Carolina at an elevation of 3700 feet (1126 m). This location, not shown on Figure 1, is only 47.5 minutes north of my study area. At Banner-Elk the annual snowfall has averaged 37 inches (940 mm) between 1951 and 1960. Since 1951 the snowiest winter there was 1960, with 91 inches (2311 mm). Almost certainly the broad upper tableland just to the south of Black Balsam Knob (Figure 1) at 5700 feet (1735 m) receives more snow than Banner-Elk which is 2000 feet (609 m) lower.

One may briefly review the snowfall data for a truly alpine area in the northern Appalachians to note similarities and bold contrasts. At Mount Washington, 6288 feet (1915 m) in New Hampshire, at latitude 44°15' north, the precipitation is close to 74 inches per year (1879 mm). R. P. Goldthwait notes (1970, page 1), "much of the precipitation comes as snow which measures 144 to 344 inches (3657-8737 mm) annual total; 203 inches (5156 mm) was the average for the first 35 years of record." Thus, there is a great contrast in snowfall between these two mountain locations which are separated by 10 degrees of latitude yet the altitudes are the same and the total annual precipitation values are the same. The mean annual temperature on Mount Washington is 27°F (-2.7°C) therefore it is obvious why this station receives more snowfall.

RAVINES OR CIRQUE-LIKE FORMS

Two small semicircular basins are present on the east side of Black Balsam Knob. One is only about the size of a large nivation hollow; the other is 600 feet deep (182 m) and topographically resembles an incipient cirque with steep bedrock walls and a long boulder talus accumulation down the middle of the ravine. The upper parts of both these "basins" have slope angles of 40 degrees. Both face eastwards and are in a very favorable position to receive wind-blown snow from a broad flat upland just to the west.

Numerous fine striations were found on quartz veins which intrude the gneiss on Black Balsam Knob. No striations have been observed on the gneiss which is weathered as much as 4 to 6 inches (101-152 mm) below the level of the quartz intrusives. The striations coincide with the direction of orientation of the cirque-like basins. These striae are probably slickensides caused by local faulting. The largest east-facing ravine on Black Balsam Knob is at an altitude of 5600 feet (1704 m). If glacial ice developed here it would have been confined to the very upper portion of the ravine, judging from its small size and crudely semicircular shape.

The most convincing cirque-like forms, seen thus far in the immediate vicinity of Black Balsam Knob, occur along the main scarp of what is locally called Pisgah Ridge (Figure 1) from Devil's Courthouse to Fork River Bald. These ravines face southeastward and would have received stronger insolation than the east facing locations, but their southeastward orientation would have provided more freeze-thaw cycles favoring cirque development. Immediately above and to the west of these southeastern facing ravines there extends a broad, flat surface that served as a snow accumulation field and from which westerly winds could have blown great amounts of drifted snow. One can assume that timberline was at an elevation lower than summit levels today. Michalek, 1968, contributed an excellent discussion of lowered timberlines and past timberline is briefly discussed in King, Neuman and Hadley (1968, p. 1) and in Hamilton (1961). Professor G. Michael Clark of the University of Tennessee has indicated (written communication, 1973) that the subject of past snowline and timberline is still unsettled.

The bottoms of these cirque-like forms along Pisgah Ridge range downward to an elevation of 4800 feet (1462 m) and the highest "headwalls" are at about 5800 feet (1765 m). As yet, the floors of these especially deep ravines have not been checked but have only been observed from neighboring peaks and from the Parkway.

Many similar forms occur across the high mountains from the Great Smokies to Mount Mitchell. Within the southern Appalachians, only one location is reported to have suffered the effects of alpine glaciation, namely Grandfather Mountain (Berkland and Raymond, 1973). Their field data has been questioned but should the data stand the test

Figure 2. Solifluction Turf-Hummock at 5800 ft. (1, 765 m) east side Black-Balsam Knob, North Carolina.

of time it should stimulate additional field research.

In the region still under investigation I have found no incontestable evidence that documents alpine glaciation. It is extremely difficult to separate forms produced by advanced nivational processes from those that may have developed from incipient glaciation.

Small scale mass-wasting still continues today on Black Balsam Knob where frost hummocks and incipient solifluction turf-lobes (Figure 2) have been encountered. To date, no patterned ground has been found but sorted forms have been reported on Whitetop Mountain in southwestern Virginia (Clark, 1968).

The best-developed and best-located cirque form seen thus far in the western part of this study area is on the north side of Richland Balsam (Figure 1). The headwall is at an elevation of 6200 feet (1887 m) and the break in slope, between headwall and gentle slope below, is at 5600 feet (1704 m). At the above location one finds an unusually well-developed truncated spur at 5400 feet (1643 m) suggesting the coalescence of two small cirque glaciers that may have been responsible for the modification of this ridge. The elevation of Richland Balsam, 6410 feet (1952 m) is only 278 feet (85 m) lower than the summit elevation of Mount Mitchell making it one of the highest peaks in the southern

Appalachians. Three deep confluent ravines on the east side of Richland Balsam may also have supported small ice masses, but as yet, I have no firm field data to support this inference. Certainly the high elevation and irregular cirque-like forms at the heads of the ravines on the north and east sides of Richland Balsam are strong candidates for modified incipient cirques partially filled by postglacial mass-wasted debris and trenched by present day stream action.

From field observations since 1967, I have seen numerous boulder talus accumulations, particularly on north and northeastern facing slopes of the highest peaks, from the Great Smoky Mountains to Mount Mitchell. These are not unique to the southern Appalachian mountains, but in fact are rather commonplace features and support the role played by more intensive freeze-thaw activity during late Pleistocene time. Climate had to be more rigorous to produce these large frost-shattered blocks and details of similar features are well documented in Michalek's work (1968). Within the immediate map area of Figure 1, well-developed boulder talus accumulations occur along the valley of the North Prong of Shining Creek and within the steep ravines on the eastern side of Pisgah Ridge. Such talus has been encountered at elevations as low as 4000 feet (1218 m).

CONCLUSIONS

To veterans of alpine glacial geology and geomorphology many of the features discussed above may seem too small and underdeveloped to suggest incipient glaciation. The reader must keep in mind the geographic setting of the southern Appalachians. This region was not covered by an ice sheet as was its counterpart to the north and furthermore, the climate ameliorated here much earlier than it did farther north. Postglacial conditions were warmer, more humid, and wetter so that mass-wasting, weathering and stream action have masked many of the earlier formed features discussed above.

Future evidence in support of mountain glaciation may come from the Black Mountains in Northern Carolina, where Mount Mitchell is located, and in the Plott Balsam Range west of this study area.

This short paper hopefully will serve as a status report of the writer's work in progress in this area and provide some preliminary findings that suggest strong nival processes and frost action, and perhaps incipient ice erosion in an area outside of the Grandfather Mountain area. Hopefully, it will stimulate more research from other interested investigators that will ultimately support or refute apline glaciation in the southern Appalachians.

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ORBICULAR ROCKS FROM DAVIE COUNTY,

NORTH CAROLINA PIEDMONT

By

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ABSTRACT

Dark-colored orbicules generally about 20 mm in diameter in a white quartzofeldspathic matrix occur in a limited exposure in eastern Davie County, North Carolina Piedmont. The orbicules have a single shell composed of radially oriented augite and hornblende. The matrix is mainly microcline, plagioclase, and quartz, with lesser amounts of hornblende and sphene. Four alternative hypotheses of origin are liquid immiscibility, mixing of magmas, comb-structure crystallization, and spherulitic recrystallization. Bulk composition falls in a field of liquid immiscibility extrapolated from the leucite-fayalite-silica synthetic system. The hypothesis of liquid immiscibility is favored, although the evidence is far from compelling.

INTRODUCTION

A very unusual type of orbicular rock occurs in a limited exposure in eastern Davie County, North Carolina Piedmont. Spheroidal greenish black orbicules set in a strongly contrasting white matrix give the rock a striking appearance. Watson (1904) gave a general description of the field relationships and petrography. This paper adds petrographic data, gives estimated chemical compositions, and presents several alternative hypotheses about origin of the orbicular rock. In a survey of main references on orbicular rocks (Leveson, 1966, and earlier references), I could find no descriptions of closely similar types. This occurrence of orbicular rock and the hypotheses concerning origin have so many obscure aspects that the explanations given here are necessarily very speculative. Nomenclature used here follows the standard terminology for orbicular rocks suggested by Leveson (1966).

Acknowledgments

William A. White of the University of North Carolina at Chapel

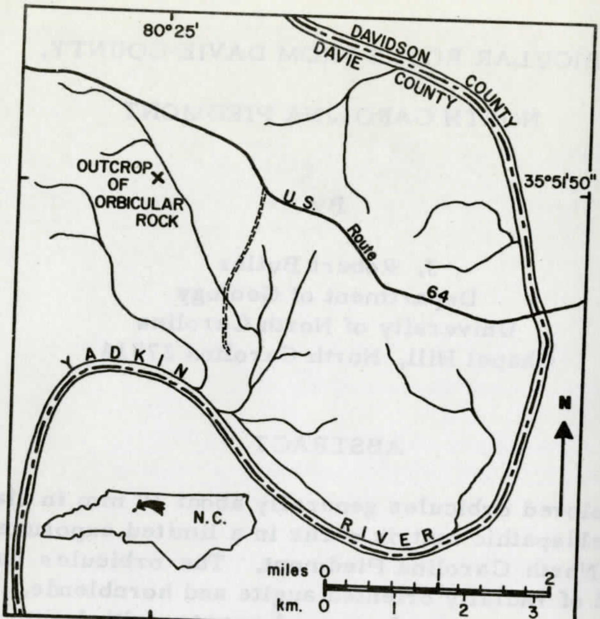

Figure 1. Index map showing location of the Davie County orbicular rock. Base map from the Churchland 7 1/2-minute quadrangle, N. C.

Hill first brought the rock to my attention. Mr. and Mrs. Peter W. Hairston kindly permitted access to their property. Expenses were covered by a grant from the University Research Council, University of North Carolina at Chapel Hill. I am grateful to the following people for suggesting improvements in the manuscript: David J. Leveson of Brooklyn College, Loren A. Raymond of Appalachian State University, and William H. Spence of North Carolina State University.

FIELD RELATIONSHIPS

The orbicular rock occurs on a small hill south of U. S. Highway 64 in eastern Davie County, North Carolina Piedmont (Figure 1). Outcrops and residual boulders (Figure 2) extend over an area about 40 m (132 ft.) long by 25 m (82 ft.) wide. The locality was probably first reported by Lewis (1893, p. 91) and was subsequently described by Watson (1904) and Watson and Laney (1906). Watson (1904, p. 296) observed that saprolite of orbicular rocks "averages several hundred feet in width and extends approximately one-half to three-quarters of a mile southwest from the knoll." He also noted sharp contacts with the country rock at several places and a loss of orbicular structure to the

Figure 2. Main outcrop area of orbicular rock.

southwest. The area is now heavily wooded, and Watson's observations could not be verified during the present study.

The orbicular rock is cut by several white aplite and pegmatite dikes and by quartz and quartz-epidote veins. These dikes are mainly less than 15 cm thick, although one saprolite exposure has pegmatitic material up to 70 cm thick. The veins range up to 5 cm in thickness.

Country rock in the vicinity of the orbicular rock, and probably extending for several kilometers on all sides, is a medium- to coarse-grained porphyritic biotite granodiorite or adamellite. This granitic rock is part of the Churchland pluton of probable middle or late Paleozoic age (Butler and Ragland, 1969). Diabase dikes (Triassic or Jurassic age) cut the pluton.

Scantly available evidence suggests that the present exposures of orbicular rock represent a small fraction of a dike or dike-like body extending northeast-southwest; however, the exposure could represent a large xenolith or roof pendant.

LITHOLOGY AND PETROGRAPHY

Nature and Size of Orbicules

The most common type of orbicular rock (Figure 3) has greenish-black, spheroidal orbicules (diameters of about 20 $\pm$ 10 mm) in a white matrix composed of plagioclase, microcline and quartz speckled with elongate hornblende crystals. Scattered brown sphene crystals are commonly visible in hand specimen. The orbicules larger than about 16 mm typically have distinct shells and cores (Figures 3, 4). The

Figure 3. Sawed slab of orbicular rock. Bar is 20 mm long.

shells are composed mainly of radially oriented augite and hornblende grains, and range from about 5 to 8 mm thick, no matter what the size of the orbicule.

Only a small fraction of the orbicules have diameters greater than 30 mm. The largest observed orbicule is 80 mm in longest dimension and has a flattened appearance.

Cores of orbicules show a wide variation in mineralogy and are generally finer grained than shells or matrix. The most common type of core has epidote, augite and sphene as major constituents. Cores of largest orbicules also contain microcline, plagioclase, quartz, and hornblende. The mineral grains are commonly equant. Epidote (pleochroic from neutral to grayish yellow) is a major constituent of most cores, but is relatively rare in rims and matrix.

Matrix

The matrix is composed mainly of medium- to coarse-grained plagioclase (mainly andesine), microcline, and quartz. Scattered prisms of greenish black hornblende up to 15 mm long and crystals of brown sphene up to 3 mm long are conspicuous in the matrix. Thin sections show that plagioclase is extensively replaced by white mica and clinozoisite. Locally microcline and quartz occur as granophyric intergrowths. Apatite is present as blunt rods up to 1.5 mm long. Minor amounts of chlorite are present as alteration products.

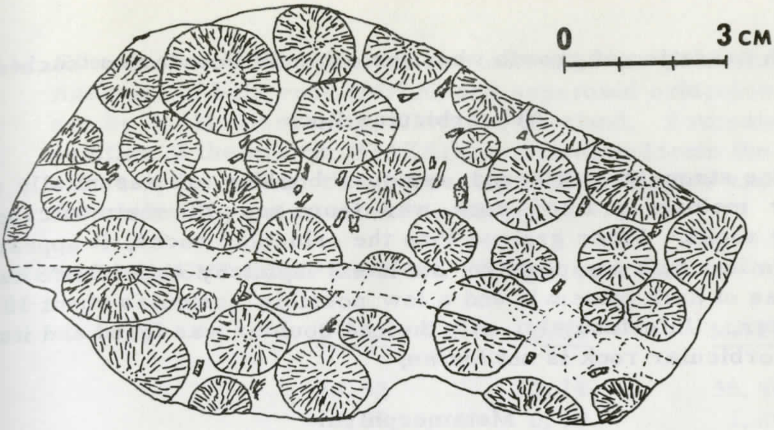

Figure 4. Tracing from sawed slab showing quartzo-feldspathic veinlet cutting orbicules.

In one slab a white veinlet 5 mm thick and at least 15 cm long cuts across the orbicules and merges with the matrix between orbicules (Figure 4). The veinlet contains approximately equal amounts of plagioclase, microcline, and quartz with an average grain size of about 0.7 mm, and has minor amounts of epidote and sphene. Separated pieces of several orbicules can be matched across the veinlet, demonstrating that it formed by dilation rather than replacement. Some plagioclase and microcline crystals extend entirely across the veinlet and partly enclose orbicule fragments, thus forming part of the matrix as well as part of the veinlet. The relationships indicate that the veinlet follows a fracture developed after the orbicules had formed.

Percentages of Orbicules and Matrix

The percentages of orbicules and matrix were determined by point-counting a group of slabs with a total area of 1550 cm^2 (240 in^2). A square grid with an interval of 1.25 cm (about 0.5 in) on clear plastic was superimposed on the slabs and 1000 points were counted. The point count indicates that orbicules make up 64.2 percent and the matrix 35.8 percent of the rock by volume in the available slabs. Subtotals for individual large slabs showed little variation of percentages, which suggests that the percentages remain nearly constant.

The percentages of orbicules and their spatial relationships suggest that the rock represents an accumulation of spheroids that settled through a fluid. Fraser (1935) found that in a compacted wet marine sand, grains made up 64.96 percent of the aggregate and pore space was 35.04 percent. The interpretation made here is that semi-rigid orbicules settled through a fluid which later crystallized to form the matrix. The orbicules would be analogous to Fraser's sand grains and the matrix to pore space. An alternate hypothesis is that orbicules grew <u>in</u>

situ, with cessation of growth when the orbicule boundaries touched.

Non-orbicular Rock

One strongly weathered residual boulder of essentially non-orbicular medium-grained rock was found near the main outcrops of orbicular rocks. Finer grained than the orbicular rocks, it appears to have a similar bulk composition but has a splotchy appearance due to aggregates of dark minerals and a few possible orbicules about 10 mm in diameter. Unfortunately, only the one boulder was found and its relation to orbicular rock is not known.

Metamorphism

The rock has been metamorphosed to some degree. Augite in orbicule shells has a clouded appearance caused by submicroscopic inclusions, which led Watson (1904) to call the mineral "diallage." On the other hand, clear fresh-looking augite of some orbicule cores is commonly intergrown with pleochroic iron-rich epidote, which otherwise is scarce in the rock. Some of the hornblende is partly altered to a pleochroic pale-green amphibole, probably actinolite.

The interpretation here is that the augite, hornblende, microcline, plagioclase, quartz, sphene, and apatite are primary igneous minerals and the granophyric intergrowths are evidence of igneous origin. Metamorphism, however, caused partial recrystallization of the rock, saussuritization of plagioclase, clouding of augite, and formation of epidote, actinolite(?), and chlorite. It is not known whether the metamorphism is contact or regional.

CHEMICAL COMPOSITION

A chemical analysis of the bulk rock (Table 1) was obtained from a split of a crushed thin slab (about 5 mm thick), cut in a random direction from a large sample (about 20 cm across) that was judged to be typical of the exposure. In thin slabs, the matrix can be broken from the orbicules. A composite orbicule sample was prepared and analyzed (Table 1). The matrix appears heterogeneous in stained slabs, so the matrix composition was calculated, using the percentages of orbicules and matrix determined by point count (after conversion from volume percent to weight percent).

The bulk composition is not closely similar to any common igneous rock (Nockolds, 1954). The orbicule composition is closest to the pyroxenite group, and matrix composition to mangerites and doreites (Nockolds, 1954, Table 5), which are intermediate to diorite and monzonite.

Trace-element data (Table 2) indicate that the orbicules are

Table 1. Chemical and Normative Composition of the Orbicular Rock. Analyses of bulk rock (BULK) and separated orbicules (ORBS) are by W. H. Herdsman, Glasgow, Scotland. Estimated composition of the matrix (MATRIX) is calculated from the chemical analyses using the percentages of orbicules and matrix determined by point-counts on slabs.

A. MAJOR-ELEMENT COMPOSITION. Weight percent.

	BULK	ORBS	MATRIX
SiO_2	53.52	52.14	56.98
TiO_2	0.74	0.49	1.29
Al_2O_3	7.52	3.23	16.80
Fe_2O_3	2.58	2.97	1.77
FeO	4.55	5.37	2.84
MnO	0.15	0.17	0.13
MgO	12.09	15.54	4.80
CaO	14.14	17.92	6.17
Na_2O	1.60	0.74	3.46
K_2O	1.15	0.24	3.11
H_2O^+	0.93	0.64	1.56
H_2O^-	0.54	0.10	0.32
P_2O_5	0.29	0.07	0.77
CO_2	0.13	0.24	0.00
	99.93	99.86	100.00

B. C.I.P.W. NORMS.

quartz	2.7	0.8	6.9
orthoclase	6.8	1.4	18.4
albite	13.5	6.3	29.3
anorthite	10.0	4.9	21.1
wollastonite	24.0	34.3	1.8
enstatite	30.1	38.7	11.9
ferrosilite	5.2	6.9	1.9
magnetite	3.7	4.3	2.6
ilmenite	1.4	0.9	2.4
apatite	0.7	0.2	1.8
calcite	0.3	0.5	0.0

relatively enriched in Cr, Cu, Li, Ni, and Sc, while the matrix is enriched in Ba, Co, Rb, Sr and Zr. These data should be used with caution because of possible metamorphic redistribution and the errors involved in separation of orbicules and calculation of matrix composition.

Table 2. Trace-element Composition. Values for bulk rock and orbicules determined by emission spectrography. Analyst: Oiva Joensuu, Miami, Florida. Values for matrix calculated as explained in Table 1. Parts per million.

	BULK	ORBS	MATRIX
Ba	650	60	1900
Be	2	3	0
Co	160	90	300
Cr	1400	2300	0
Cu	25	28	19
La	22	<10	-
Li	15	20	5
Ni	230	280	120
Rb	19	2	56
Sc	45	65	3
Sr	1100	370	2650
V	160	160	160
Y	22	<10	-
Zr	40	20	80

HYPOTHESES OF ORIGIN

General

The orbicular rock must have formed by an extremely rare combination of processes. I can find no descriptions of closely similar orbicular rocks. Leveson (1966) pointed out that orbicule formation hypotheses include both magmatic and metamorphic origins, and that no single hypothesis provides a general explanation.

In this section, four basically different hypotheses are considered for the Davie County orbicular rock: (1) Liquid immiscibility, (2) Mixing of magmas, (3) Comb-structure crystallization, (4) Spherulitic recrystallization. Discussion of these hypotheses is very speculative and possibly none is correct.

Liquid Immiscibility

A magma of unusual composition may separate into two immiscible parts, with mafic liquid forming spherical drops in a more felsic magma. Crystallization of pyroxene and amphibole begins first in the mafic portion; these crystals form radial rims that become relatively rigid. The armored mafic spheroids settle through the felsic liquid and accumulate into a compact mass with "pore space" similar in percentage to that observed in sediments. Crystallization continues, with some chemical interchange by diffusion between orbicules and matrix. As crystallization nears completion, some fractures form that cut across solid orbicules and are filled with still-liquid felsic constituents. At some time after the mass crystallizes, metamorphism of low to medium rank causes some recrystallization and replacement of original minerals.

Greig (1927) showed that fields of liquid immiscibility observed in synthetic systems were seen only in silica-rich liquids unlike any common types of igneous rocks. Roedder (1951) located a previously unknown field of liquid immiscibility in the system silica-leucite-fayalite, a part of the system SiO_2-K_2O-FeO-Al_2O_3. Figure 5, the type of diagram used by Greig, shows fields of liquid immiscibility observed in simple systems studied by Roedder and earlier workers. The diagrams and data were converted from weight per cent to molecular per cent. Use of the diagram is based on the possibly incorrect assumption that addition of Na_2O, CaO, MgO, and minor elements to Roedder's SiO_2-K_2O-FeO-Al_2O_3 system would not greatly affect the fields of liquid immiscibility. Roedder (1956) cited unpublished experimental data indicating that addition of some other components did eliminate the immiscibility field within the diagram. On the other hand, Roedder and Weiblen (1970) found evidence for silicate liquid immiscibility in lunar magmas. Compositions of the two lunar liquids did not differ greatly from certain immiscible liquid pairs in the synthetic system leucite-fayalite-silica. Bulk composition of the orbicular rock plots within the field of two liquids (Figure 5), while the orbicule composition plots on the edge of the field and the matrix composition plots along a general trend defined by common igneous rocks. This diagram lends some support to the hypothesis of liquid immiscibility, although large extrapolations of experimental data are involved. Even if valid, this concept would apply only to a very limited range of rock compositions and physical conditions.

Mixing of Magmas

Mafic globules in a felsic magma may originate by mixing of magmas, with subsequent history being similar to that described above. This model requires a mechanism of dispersing the mafic magma so that it becomes discrete globules. Such a mechanism might be provided by forcefully, even explosively, injecting mafic magma through a narrow conduit into a pool of felsic magma. There would be an exchange of heat between the supposedly much hotter mafic magma and cooler felsic material. The result would be rapid crystallization of the mafic part and heating of the felsic part. Chemical interchange between orbicules and matrix and recrystallization of the rock may have modified the original composition and grain size.

Walker and Skelhorn (1966) reviewed numerous occurrences of two magmas with different compositions that were mixed, but retained

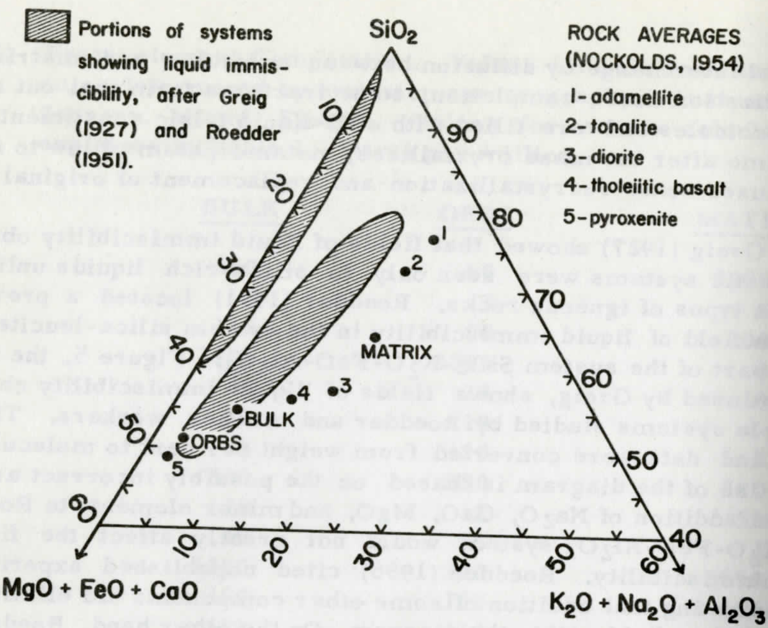

Figure 5. Part of triangular diagram for combined oxides, showing fields of liquid immiscibility in simpler systems and analyses of selected rocks. Molecular percent. ORBS-separated orbicules, BULK-bulk orbicular rock, MATRIX-calculated composition for matrix of orbicular rock.

their identities as separate phases. None of their examples is physically similar to the Davie County orbicular rock; however, it appears that mechanisms for dispersing droplets of one magma in another may exist. Walker (1962) observed bubbles of basaltic glass dispersed through a felsic welded tuff in Iceland. The basaltic bubbles, normally less than one-half mm in diameter, are commonly spheroidal in shape (p. 284) and make up about two percent of the unit.

Conclusions of Yoder (1973) regarding mixing of rhyolitic and basaltic magmas may also have bearing on this hypothesis. He found that such magmas in contact would maintain sharp contacts over short periods. The behavior was not due to liquid immiscibility. Yoder (1973, Fig. 9) shows both experimental and natural contacts of rhyolitic and basaltic glasses. The contacts of the mafic part are lobate toward the felsic and what appear to be rounded globules of mafic material occur in the felsic. It seems possible that forceful injection of thin streams of mafic magma into a magma chamber of felsic material could result in mafic globules that would retain their identity through cooling and crystallization, but it is difficult to imagine production of such numerous and relatively uniform-sized globules as in the orbicular rock.

Comb-structure Crystallization

The term comb layering describes a type of layering in granitoid rocks in which constituent crystals are oriented approximately perpendicular to planes of layering (Moore and Lockwood, 1973). Orbicules with radial crystals are associated with comb layering at many localities in the Mesozoic batholiths of California. According to Moore and Lockwood (1973, p. 13), "orbs characteristically have multiple layers, with nuclei of differing rock types" and "the nuclei vary greatly in shape and range in size from a few centimeters to several decimeters." Comb layering and orbicules were probably formed by "aqueous fluids that migrated upward along contacts between magma and wall-rock or along the interface between magma and previously solidified melt" (p. 1). Perhaps the Davie County orbicular rock formed by processes similar to those postulated by Moore and Lockwood, here called comb-structure crystallization to include formation of both comb layering and associated orbicules. This hypothesis is attractive, because it apparently can explain orbicule-filled dikes and pipes (Moore and Lockwood, 1973, p. 17-18 and Fig. 17). On the other hand, the Davie County orbicules have neither multiple shells nor nuclei of differing rock types such as described in the California localities. Also, individual crystals at the Davie County locality do not show the branching tree-like habit characteristic of comb layering.

Spherulitic Recrystallization

The fourth hypothesis involves recrystallization in a manner somewhat similar to formation of spherulites. Spherulites occur mainly in felsic volcanic rocks, and range in size from microscopic pellets to spheres more than three meters in diameter (Johannsen, 1939, p. 15-16). They seem to form best under conditions of rapid crystallization. There are several reasons why this hypothesis is the least likely. The surrounding pluton is granodiorite and adamellite, and probably was not hot or dry enough to form augite by contact metamorphism. Spherulitic crystallization generally forms structures not drastically different in composition from the matrix. Patterns of spherulites are different from the orbicular rock, at least in the illustrations that I can find in various references and in several thin sections. For example, spherulites in many cases have nearly planar mutual boundaries and may be flat-sided or nearly polygonal, which was not observed in the orbicular rock.

CONCLUSIONS

In my opinion, the best explanation for the orbicular rock involves liquid immiscibility in a magma of very unusual composition.

Critical parts of the puzzle are missing, so this conclusion is tentative. A mafic part of the magma separated into globules, which became armored by radial crystallization of shells composed of augite and hornblende. These globules settled through the felsic fraction to accumulate as a mass of spheroids with interstitial liquid. There was probably some chemical interaction between orbicules and matrix during crystallization. Later metamorphism caused some mineralogical changes, but very little chemical rearrangement.

The origin of a magma with such unusual composition poses a problem. A mixture of about 60 percent Mg-rich augite and 40 percent granodiorite gives a bulk composition similar to the orbicular rock. Since the orbicules may represent gravitational concentration of mafic constituents, the original magma may have been much closer in composition to granodiorite. Perhaps a superheated granodiorite magma had its composition altered by reacting with and resorbing xenocrysts of mafic minerals or pyroxenite xenoliths, then separated into two liquids upon further cooling.

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POSSIBLE INFLUENCES OF SUBPEAT TOPOGRAPHY AND SEDIMENT TYPE UPON THE DEVELOPMENT OF THE OKEFENOKEE SWAMP-MARSH COMPLEX OF GEORGIA

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ABSTRACT

Forty-three piston cores and numerous depth probings were taken in the Okefenokee Swamp of Georgia. The purpose of this sampling was to obtain some accurate measurements of peat thicknesses in this area and to determine the composition of the substrate upon which the peat had developed.

The peat was found to range in thickness from 1 to 12 1/2 feet. The positions of the large marshes (i. e., Floyd's, Chase, and Grand Prairies) were found to be coincident with the deepest deposits. Previous paleobotanical investigations have revealed that the vegetational environments in these prairies have remained nearly constant throughout most of their history. It is therefore suggested that (in addition to fires) the development and persistence of these large marshes is directly related to the positions of subpeat topographic depressions. Furthermore, the positions of other surface features (such as Big Water Lake and Minnie's lake) also seem to be related to the prepeat topography.

Much of the peat-forming area was found to be underlain by a relatively pure sand; however, sandy clays and nearly pure clays were also found and tended to be more common in deeper portions of the Swamp. These clays could have played an important part in increasing the hydroperiods in these areas.

The Okefenokee Swamp is developed upon a Pleistocene (Interglacial) marine terrace which is situated about 100 to 160 feet above present sea level. However, radiocarbon dating of basal peats from cores taken within the Swamp suggests that the Swamp did not start forming on this surface (or at least peat did not begin to accumulate) until about 6500 yrs. B. P. Thus, the upper marine (Pleistocene) sediments of the terrace were probably exposed to subaerial leaching and reworking for a considerable length of time prior to the onset of modern peat sedimentation. In addition, it is likely that terrestrial geomorphic patterns could have been superimposed upon the marine terrace by

streams, winds, or other surface processes during this hiatus.

At the beginning of peat deposition, the Okefenokee probably consisted of a series of disconnected or partially connected open fresh water marshes occurring in about the same positions as the present-day Okefenokee prairies. These marshes contained water-lilies, sedges, and other herbaceous aquatics (similar to those found in the modern-day prairies and "boat runs") with some cypresses and other swamp trees at their margins and pines and hardwoods in the more-sandy, better-drained upland regions. A more accurate representation of the paleogeography of the primordial Okefenokee (i. e., whether it originated in the lows of a dendritic stream system, or as a series of lakes, or both) awaits further investigation.

INTRODUCTION

An investigation of the peat sediments in the Okefenokee Swamp of Georgia was begun in the summer of 1970 and is presently continuing. The overall goals of this project are: (1) to establish the paleobotanical and mineralogical composition of the peat sediments in the Swamp and to correlate these sediments with modern environments of deposition; (2) to determine the vertical and horizontal distribution of peat types, the geometry of the peat deposits, and the composition of the substrate and to use this information to help reconstruct the recent geologic history of this region; and (3) to study the alteration of various plant fragments with depth and age, and to determine the potentials of these tissues to produce the macerals found in ancient bituminous coals. This paper reports on only one aspect of the recent geologic history of this region; i. e., the possible implications of subpeat topography and sediment types upon the development of peat, and upon the occurrence and distribution of modern surface features.

The Okefenokee Swamp is a paludal region of over 600 sq. miles located primarily in southern Georgia, but crossing slightly into northern Florida (Figure 1). Most of the swamp is now contained within the Okefenokee National Wildlife Refuge. The Okefenokee is not really a swamp in the classic sense, but is better described as a swamp-marsh complex. Large open marshes (i. e., relatively unforested wetlands) may make up as much as 25% of the region. These marshes are known locally as prairies (e. g., Grand Prairie and Chase Prairie, Figure 1). The prairies are dominated by herbaceous aquatic plants such as, the white water lily (Nymphaea odorata Ait.), bladderworts (Utricularia spp.), and maidencane (Panicum hemitomon Schultes); whereas, the swamps are dominated by tree vegetation, such as, cypresses (especially, Taxodium ascendens Brogn.), gums (Nyssa spp.), and bays. Sand islands covered with pines, hardwoods, and other upland plants are situated within this paludal region.

The Okefenokee Swamp is developed upon the eastern margin of

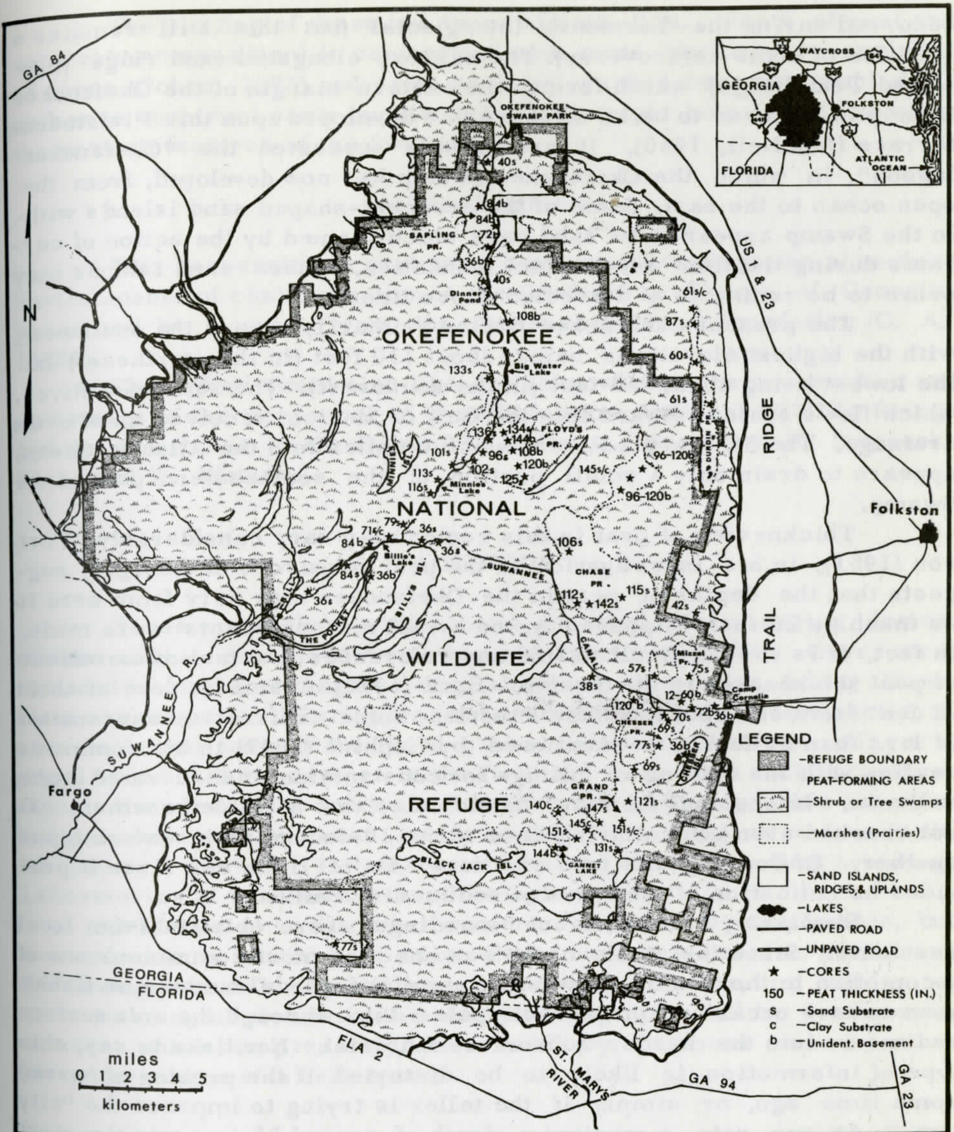

Figure 1. Index map of coring locations.

a gently-sloping, marine terrace situated upon the eastern part of the Atlantic Coastal Plain. The terrace is constructed on Pleistocene sediments at an altitude of 100 to 160 feet above sea level (MacNeil, 1950). The surface was undoubtedly formed during an early or middle Pleistocene (Interglacial) period when the sea level was relatively high. Some authors (e. g., MacNeil, 1950; Vernon, 1951) believe this to have occurred during the Yarmouth Interglacial (but this still remains a question of some controversy). The narrow, elongated sand ridge (now called TrailRidge) which forms the eastern margin of the Okefenokee Swamp is believed to be an offshore bar developed upon this Pleistocene terrace (MacNeil, 1950). It presumably separated the "Okefenokee lagoon", in which the Okefenokee Swamp has now developed, from the open ocean to the east. Some of the crescent-shaped sand islands within the Swamp appear to be sand bars which formed by the action of currents during the time of marine inundation. Other sand islands may prove to be remnants of dissected shore lines.

The present Okefenokee drains primarily toward the southwest, with the highest elevations being about 130 feet (in the northeast) and the lowest being about 105 feet (in the southwest). The Suwannee River, which flows southwestward into the Gulf of Mexico, receives most of the drainage. The St. Mary's River, which empties into the Atlantic Ocean, appears to drain only a small portion of the southeastern part of the Swamp.

Thicknesses of peat in this region may vary considerably. Fortson (1961), in his investigation of the peat resources of Georgia, suggests that the depths of peat in the Okefenokee may vary from zero to as much as 20 feet. Apparently, no actual measurements were made. In fact, it is very difficult to find any references to actual measurements of peat thicknesses in the Swamp. Bond (1970) obtained a core of about 12 feet from one spot in Chase Prairie. Some indirect measurements of less than 20 feet are mentioned by Hopkins (1947) in his comments dealing with the driving of pilings for construction of an elevated logging railroad. In one unusual case, he indicates that a stable basement was not reached even after driving down three 20-foot pilings, one on top of another. Of course, this type of "data" is of little use since it provides no indication of the types of sediments involved.

Similarly, other indirect "soundings" were obtained from local residents. Since the "pole boat" was once the most common means of locomotion in the Swamp, it was not unusual for local hunters or fishermen to have occasion to push their poles down through the soft surface sediments into the harder subsurface material. Needless to say, this type of information is likely to be distorted if the probing occurred some time ago, or simply if the teller is trying to impress the "city boy." At any rate, a maximum depth of around 25 feet was the most recurring value obtained from the above source.

Because of the dearth of accurate published depth data, this author felt the need to obtain some measurements of this type. This paper thus reports the depths of over 40 cores and numerous probings taken over a large portion of the Swamp.

In addition to the lack of depth measurements, the author discovered a lack of accurate information as to the composition of the substrate upon which the peat had accumulated. This type of information was also available from the cores and is recorded herein.

The plotting of these two types of data (i.e., depth of peat and substrate composition) in conjunction with some preliminary analyses of cores (Cohen, 1972) and some new radiocarbon dates has enabled this author to construct some hypotheses which might help explain some of the surface features of the Swamp.

Acknowledgments

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METHODS

Cores were taken with a hand-operated piston-coring device similar to that described by Cohen and Spackman (1972). Cores were obtained in four-foot lengths using aluminum irrigation tubing three inches in diameter. Coring was not terminated until the last section of core contained some of the hard-packed subpeat substrate. All cores were sealed in their casings in the field and were taken back to Southern Illinois University for analysis. Descriptions, photographic records, and paleobotanical analyses (at three-inch intervals) were done in the laboratory. Carbon 14 dates were obtained from the Geochronology Laboratories of the University of Georgia.

Selection of coring sites was done as randomly as possible, but was undoubtedly biased by the fact that I was restricted to portions of the Swamp accessible by shallow-draft boats, airboats, or roads. A more meaningful picture of the subpeat topography will be forthcoming as areas which are presently inaccessible by the above means are explored by means of a helicopter.

DATA AND DISCUSSION

Peat thicknesses were found to range from 12 to 151 inches in the sampling area (Figure 1). The shallowest peats occurred near the islands and edges of the Swamp, and the deepest almost always occurred in the prairies. The forested portions of the Okefenokee were generally less than 7 feet in depth. Both Durdin and Chesser Prairies, which are relatively narrow strips of marshland, had peat depths between five and seven feet. The larger prairies; that is, Grand, Chase, and Floyd's Prairies, ranged from 8 to 12 1/2 feet in depth. Paleobotanical analyses of cores from Durdin and Chesser Prairies indicate that both of these smaller prairies were once forested (swamp) areas which have been converted to open marshes by fires (Cohen, 1972; Cohen, In Press). On the other hand, all of the large prairies were found to have been open marshes from the time of their inception, and to have remained open marshes through out most of their history (Cohen, 1972; Cohen, In Press).

It is logical (as suggested by the work of Cypert, 1961) that fires have also played an important and perhaps necessary part in perpetuating the open areas by burning off any tree vegetation which might have begun to encroach upon them. However, fires alone could not have sustained prairies in the same relative positions throughout the past 6500 years without the prior presence of topographic lows (i. e., depressions filled with deeper peat). Trees attempting to become established in deeper peat could easily be destroyed by fire since they would be almost entirely rooted in combustible material; whereas, trees growing in shallow peat would be more difficult to destroy since they would be rooted primarily in the underlying protective sands or clays.

Some indications of the ages of the basal peats which have been formed in the Okefenokee are suggested by some new Carbon 14 dates recently obtained by the author (Table 1). Note that the oldest dates occur in the three largest prairies (Floyd's, Grand, and Chase Prairies) and that all of these prairies also contained basal peats of approximately the same age (about 6500 yrs. B.P.). At this time, the sea would have been at least 34 feet lower than it is today (Neumann, 1971) and 134 feet lower than the lowest part of the Okefenokee terrace. Needless to say, all of these peats were formed in fresh water environments. In addition, these dates indicate that plenty of time was available between withdrawal of the Pleistocene sea and formation of these Holocene peats for leaching and reworking of the upper Pleistocene sediments. This would account for the lack of any marine fossils in the sands and clays directly underlying the peats and also the lack of any secondary influences of the marine deposits upon the first-formed peats (e.g., no increase in sulfur content (Cohen, 1972)).

A cross-section of the Okefenokee from Billy's Lake to the Okefenokee Swamp Park is shown in Figure 2. It is evident that the open marshes (prairies) are situated above subpeat topographic depressions. The deepest depression seems to correlate with Floyd's Prairie, and a shallower one with Sapling Prairie. The only exception to this is the depression beneath Minnie's Lake, a site which is presently occupied by a dense, mature cypress swamp. Interestingly enough, paleobotanical analyses of cores from this area (Cohen, In Manuscript) reveal that this region was formerly an open prairie which has since (for some as-yet unexplained reason) been overgrown by swamp vegetation.

Another interesting feature of this cross-section is the relatively rapid decrease in depth as one goes north from the Big Water Lake

Table 1. Location and Description of Carbon 14 Analyses of Basal Samples from the Okefenokee Swamp.

Core Location	Depth	Approx. elevation above sea level	Sediment Type	Radiocarbon* Date (yrs. B. P.)
Floyd's Prairie Canoe Trail (30°15'50" Lat.; 82°16'30" Long.)	122-124 in.	104 ft.	peaty-sand	6585±160
Gannet Lake Trail in Grand Prairie (30°39'50" Lat.; 82°14'10" Long.)	163-169 in.	111 ft.	peaty-sand	6235±105
Chesser Prairie 3B (30°42'10" Lat.; 82°11'50" Long.)	75-79 in.	112 ft.	sandy-peat	3910±85
Mizzell Prairie (30°45'40" Lat.; 82°10'10" Long.)	74-78 in.	112 ft.	peaty-sand	4385±95
Dinner Pond (30°58'40" Lat.; 82°17'0" Long.)	41 in.	116 ft.	wood	2950±160
Chase Prairie (30°48'20" Lat.; 82°14'40" Long.)	109-113 in.	110 ft.	sandy-peat	6490±80

*All analyses made by the Geochronology Laboratory of the University of Georgia, Athens, Georgia.

region to Dinner Pond. This prominent subpeat terrace-like feature perhaps represents the position of a former shoreline as the sea withdrew toward the south. The younger age of the basal peat found upon this terrace (i.e., at Dinner Pond) may suggest that as the water table rose, the swamp grew outward and upward from topographic depressions (such as that at Floyd's Prairie) and eventually covered up some of the topographically higher features of the region. It is interesting to note that this outward and upward growth from topographic depressions was found by Whitehead (1972) to be the type of development which he encountered in the Dismal Swamp of Virginia. However, because of his ability to obtain a more widespread distribution of probings (due to the presence of numerous manmade drainage canals), Whitehead was able to establish that the Dismal Swamp developed out of and eventually submerged a former dendritic stream system. A dentritic stream pattern superimposed upon a marine surface may also account for some

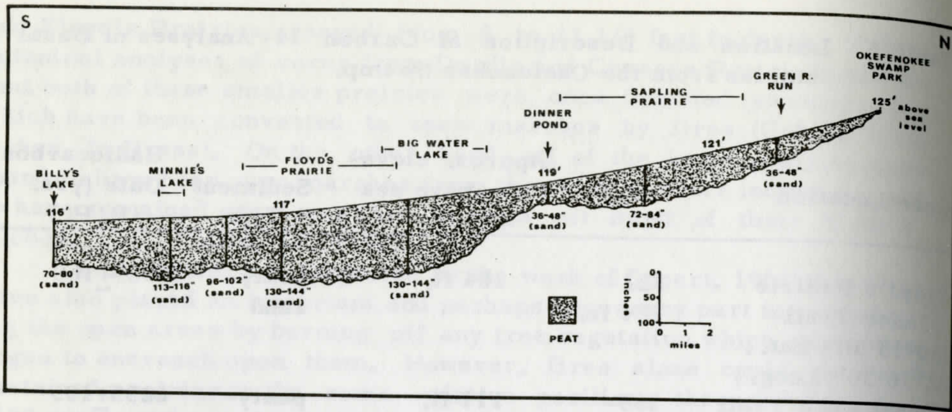

Figure 2. Schematic cross-section of the Okefenokee Swamp from Billy's Lake to the Okefenokee Swamp Park showing peat thickness and subpeat topography.

of the topographic features beneath the Okefenokee Swamp. The present geomorphology of the Swamp would suggest that if a submerged stream system does exist it would probably flow in the direction of the drainage of the present Suwannee River. This is one hypothesis which will certainly be tested as more data is obtained from areas of the Okefenokee which are presently accessible only by helicopter.

Two types of lakes can be observed in the Okefenokee. The first type is roughly circular in outline and occurs only in the prairie regions (e.g., Gannet Lake and Monkey Lake). This type probably resulted from burning of pockets of peat during severe droughts and fires (Cypert, 1961). The second type has a narrow elongated form (e.g., Billy's Lake, Minnie's Lake, and Big Water Lake), and simply appears to be a wider and deeper portion of a stream (or "boat run"). This second type of "lake" also appears to be directly related to the preswamp topography. Both Minnie's and Big Water Lakes are located just above prepeat topographic depressions (Figure 2). Both of these streams have also cut down through the peat and into the sandy substrate. Since the composition of the sediment upon which the stream is cutting does not seem to change from north to south, the presence of the "lakes" must be related to the presence of the depressions and not to any change in the petrographic character of the bed or in the sediment load carried by the stream. The elongated nature of these lakes (and, in fact, the presence of natural "boat runs" and numerous elongated cypress "strands") may be related to remnants of a prepeat stream pattern developed upon the Okefenokee terrace. It is also possible that some of the narrow prairies (such as, Durdin and Chesser Prairies) might be related to such features. However, it is somewhat more difficult to account for some of the more oval features in the Swamp. Perhaps these are inundated "Carolina Bays" superimposed upon a stream pattern; however, such ideas are purely speculative at this point.

Hearsay evidence and a few published reports (Smedley, 1968; Veatch and Stephenson, 1911) suggested that the Okefenokee was generally developed upon a relatively clean, white sand similar to that found on the islands and surrounding uplands. Most of the cores obtained for this study did penetrate down to a sand basement. However, although some of the cores ended in a light gray or tan (slightly carbonaceous), well-rounded sand, many of the cores ended in a darker gray, clayey sand; and six of the cores ended in a dense (unstratified) sticky, blue-gray clay or slightly sandy clay. These clays seem to be most common in the deeper portions of the Swamp (Figure 1). Herrick (1970) identified a low permeability, marine Pliocene clay at a depth of about 65 ft. in a test hole drilled on Trail Ridge. Since Trail Ridge is as much as 30 feet higher than the surface of the Swamp, it is possible that the clays encountered at 12 to 15 feet beneath the peat surface could be outcrops of this same Pliocene clay. However, since no marine fossils were found in these clays, it is equally possible that they are reworked sediments deposited in stream channels or lakes developed upon the Okefenokee Terrace sometime after withdrawal of the sea and before deposition of the Holocene peats. At any rate, these clays would certainly have some effect upon the permeability of the substrate and could be a factor in the retention of standing water in the Swamp.

CONCLUSIONS

The greatest depth of peat so far found in the Okefenokee Swamp is slightly over 12 1/2 feet. No depths of peat have yet been encountered which approach the mid-20 foot figure implied in the literature or conveyed to this author by local residents. It is possible that previous reports of peat approaching this depth could actually have been the combined depth of both peat and underlying clays. Since most of the previous records of sediment depths were undoubtedly obtained simply by probing with a "push pole," it is likely that the prober may have inadvertently probed in those regions in which the peat was underlain by wet clay. This author has cored very easily in such areas to a depth of about 15 feet and could have cored further but for the fact that his interests lay mainly in the overlying peat.

The greatest depths of peat have been found in the prairies (marshlands). In other words, the major present-day marshes (i.e., Grand Prairie, Floyd's Prairie, and Chase Prairie) are situated above prepeat topographic lows. It is very unlikely that this relationship could be coincidental. It is therefore suggested that this association is a genetic one.

The position of the wide places in the north-south boat run from Dinner Pond to Suwannee River (i.e., the formation of Big Water "Lake" and Minnie's "Lake") also appear to be coincident with depressions in the subpeat topographic surface.

The substrate upon which the primordial Okefenokee developed was originally said to be a relatively pure sand similar to that found in the islands and uplands. However, in several locations (especially in the deeper parts of the Swamp), bluish-gray sticky clay or sandy clay was found. These clays may play a part in reducing the permeability and lengthening the hydroperiod in some parts of the Swamp.

Basal peats found in the deepest parts of the Swamp (i.e., in the large prairies) are formed of debris derived from water lilies and other fresh-water, herbaceous aquatics (Cohen, 1972). These peat types are nearly identical to those which are presently forming in the modern Okefenokee prairies. Thus, neither the environments of deposition nor the relative positions of the large prairies have greatly changed from the beginning of peat formation in this area. This is not to suggest that some change in size and shape of these marshlands could not have occurred.

Although the Okefenokee Swamp is developed on a marine terrace, no brackish or marine microfossils or halophytic plant remains have been found in the basal Okefenokee peats or in the immediately underlying sandy or clayey substrate. The reasons for this became obvious after obtaining C^{14} dates for basal peat sediments from within the Swamp. The oldest peats so far encountered in this region were only about 6500 yrs. B. P. Thus, the earliest peat formations occurred a considerable length of time after withdrawal of the mid-Pleistocene sea. The sandy and clayey lagoonal sediments upon which the Okefenokee developed were probably thoroughly leached and reworked before this time.

The fact that the oldest peats can be found at the base of the deepest deposits (i. e., in subpeat topographic lows) indicates (along with other evidence above) that the primordial Okefenokee "Swamp" was a series of disconnected or perhaps partially connected depressions occupied by fresh-water marsh vegetation. As peat accumulated and the water table rose, the swamp grew vertically and laterally until it eventually submerged some topographically higher features. Whether any of these topographic depressions on the prepeat surface represent submerged stream channels (as is the case in the Dismal Swamp (Whitehead, 1972)) is a hypothesis which will be tested as more data becomes available from less accessible portions of this region.

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GEOLOGY OF THE SMOKE HOLE REGION OF WEST VIRGINIA

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ABSTRACT

A detailed, geologic study of the Smoke Hole Region, Grant and Pendleton Counties, West Virginia, shows the dominant structural feature to be the Cave Mountain anticline. Throughout the region beds of the Juniata Formation of Upper Ordovician age through the Marcellus Shales of lower Middle Devonian age are exposed. The Cave Mountain anticline is an asymmetrical anticline slightly overturned to the northwest. The anticline generally trends N34°E and is a doubly-plunging fold showing a clumination within the central part of the Smoke Hole Region. The Cave Mountain anticline consists of subparallel, southeast-dipping reverse faults having a maximum stratigraphic throw of 1900 feet and northwest-striking, strike-slip, cross-faults with a maximum net-slip of 100 feet. The reverse faults are considered high-angle thrust faults representing listric surfaces due to the curvature of the fault planes with depth. Curvature of the Cave Mountain anticline is attributed to the varied forward motions of the thrusted blocks. Subsidiary folding and the thrusting show an average of 54 percent compression of the Ordovician through Devonian stratigraphic sequence. Chi square tests indicate that the thrust faults are the dominant factors influencing the drainage pattern of the South Branch River.

INTRODUCTION

The Smoke Hole Region (locally known as the Smoke Holes) represents approximately 40 square miles (100 square kilometers) of the Spruce Knob-Seneca Rocks National Recreational Area, within the Monongahela National Forest of West Virginia (Figure 1). The study area is bordered on the west by the westernmost large structure of the Appalachian Valley and Ridge Province, the Wills Mountain anticline. The Smoke Holes are located on the Onego and Petersburg 15-minute topographic maps (1926), between lat 38° 45' 40" and 39° 00' 00" N., and long 79° 20' 00" and 79° 10' 00" W. Access to the Smoke Holes is U. S. Forest Service Road 74, which traverses the entire length of the

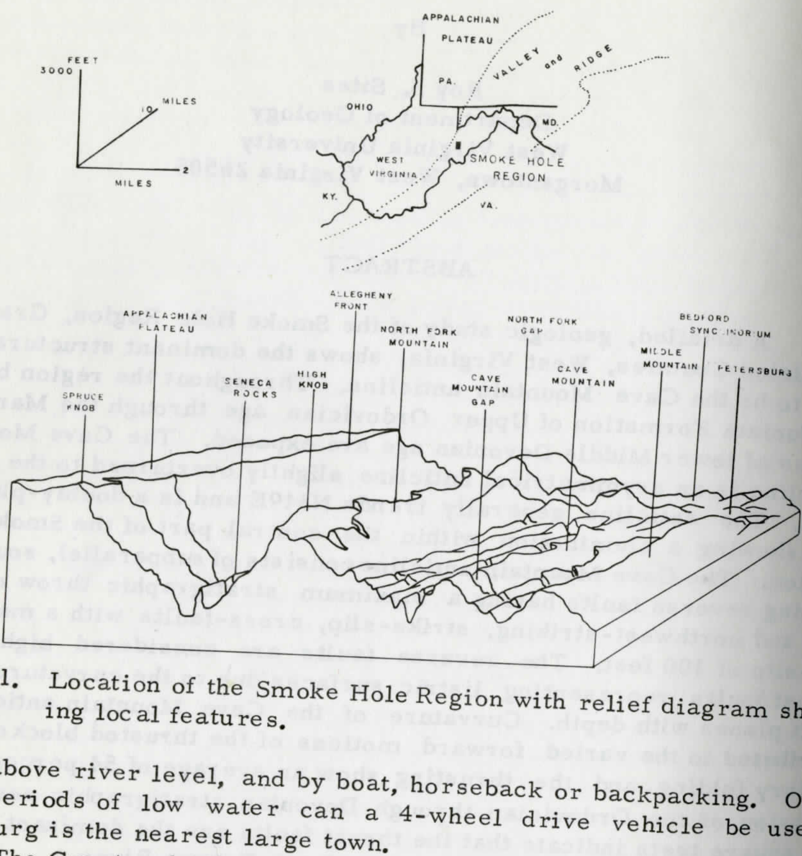

Figure 1. Location of the Smoke Hole Region with relief diagram showing local features.

region above river level, and by boat, horseback or backpacking. Only during periods of low water can a 4-wheel drive vehicle be used. Petersburg is the nearest large town.

The Grant and Mineral County Geologic Report (Reger, 1923 and Reger and Tucker, 1924) and the Pendleton County Geologic Report (Tilton, et al, 1927) contain the only geologic maps that included the Smoke Holes in detail until the present study.

Clark (1967), completed a dissertation concerning the structural geomorphology of part of the Wills Mountain anticline north of, and including, North Fork Gap. Clark's evaluation of the various hypotheses on the origin of water gaps located within the area led the present author to present new criteria for a reinterpretation of this area (Sites, 1970). Perry's (1971) dissertation on the structural development of the Nittany anticlinorium in Pendleton County is a recent regional study of the area to the west and south. Detailed geological mapping on a scale of 1:15,600 by the author and construction of cross-sections and longitudinal-sections show several previously-unmapped folds and faults (Sites, 1971). Paleozoic sedimentary rocks, ranging in age from the Upper Ordovician through the Middle Devonian, comprise the stratigraphic sequence exposed in the Smoke Holes (Figure 2).

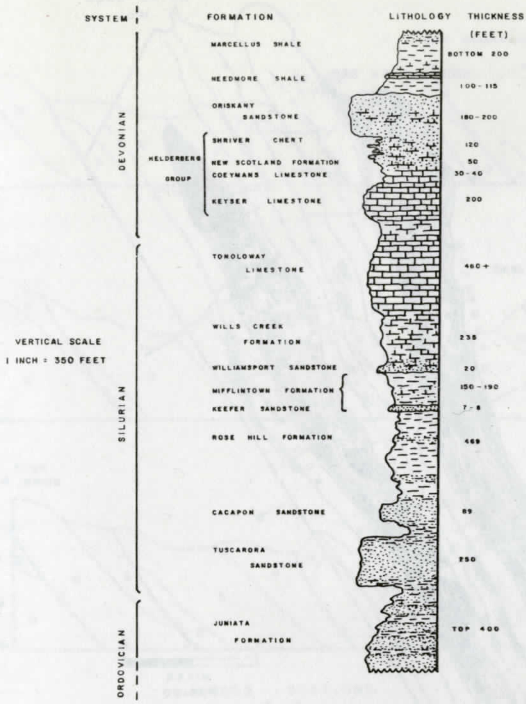

Figure 2. Stratigraphic section of the Smoke Hole Region, West Virginia.

Acknowledgments

The writer acknowledges the suggestions given by Dana Wells, Alan Donaldson and John Renton of West Virginia University. Gratefully appreciated is the field assistance given by Russell Wheeler, also of West Virginia University. The assistance of Thomas Hubbard and Lynn McClure of the U. S. Forest Service, Potomac District, Monongahela National Forest, and the assistance and suggestions of Robert Erwin of the West Virginia Geologic and Economic Survey are appreciated. Acknowledgments are also expressed to the considerate land owners throughout the Smoke Holes for their cooperation.

PHYSIOGRAPHY

North Fork Mountain and Middle Mountain are the two prominent topographic features that form, respectively, the western and eastern boundaries of the Smoke Holes (Figure 1). Along the northwestern boundary of the Smoke Holes is North Fork Gap, the major water gap

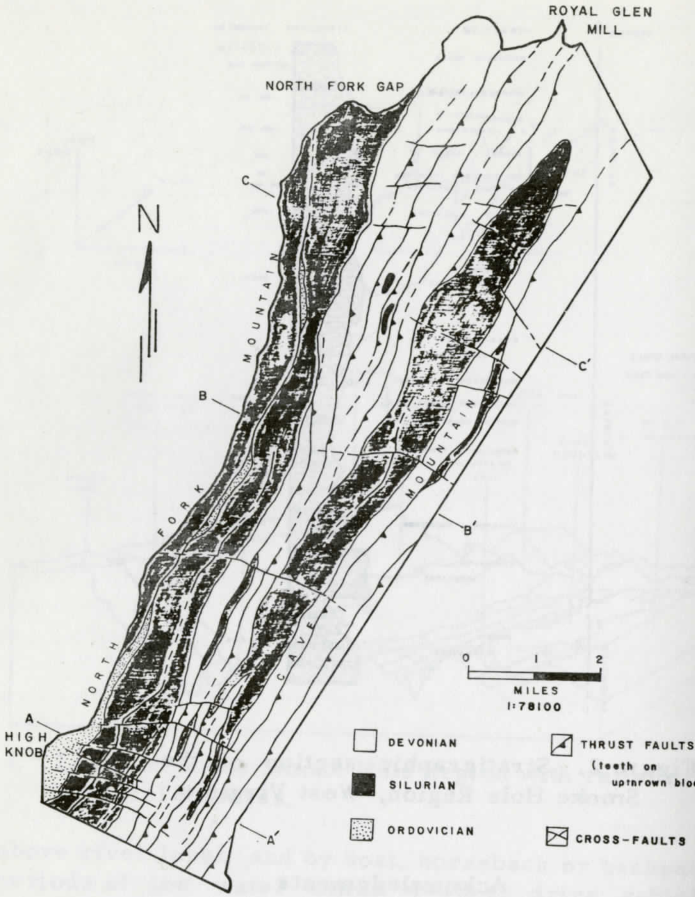

Figure 3. Generalized geologic map of the Smoke Hole Region. Doubly plunging Cave Mountain anticline with associated faulting is shown.

through North Fork Mountain. North Fork Mountain is the eastern limb of the Wills Mountain anticline, is a continuous anticlinal ridge through this region of West Virginia, reaches an altitude of 3825 feet, shows a maximum topographic relief of 2258 feet, and is capped by the Tuscarora Sandstone of Silurian age except at High Knob where the Juniata Formation of Ordovician age has been reverse faulted over the Tuscarora Sandstone (Figures 3 and 4). Middle Mountain is a continuous synclinal ridge, reaches an altitude of 2643 feet, shows a maximum topographic relief of 1593 feet and is capped by the Chemung Formation of Devonian age.

The South Branch of the Potomac River flows through a narrow gorge for the entire length of the Smoke Holes, entering the southern

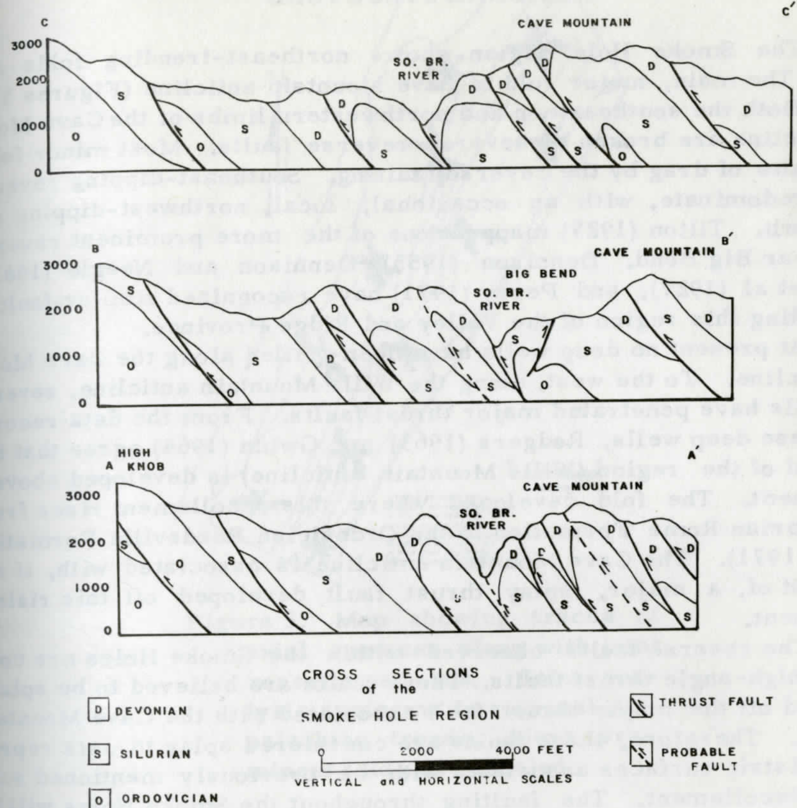

Figure 4. Geologic cross-sections showing typical structure observed within the Smoke Hole Region, West Virginia. Notice the associated folding with faulting. Overturning and chevron folds are observed through the sections.

end via the major water gaps developed through Little Mountain and Cave Mountain (Figure 1). The only water gaps through the Cave Mountain anticline are located near each end of the Smoke Holes. The river meanders through the gorge and emerges at the extreme northern end near Petersburg. Cave Mountain is the main topographic and structural feature within the Smoke Holes. It reaches an altitude of 2825 feet, shows a maximum topographic relief of 1675 feet, and is capped by the Tonoloway Limestone of Silurian age and the Helderberg Limestone of Lower Devonian age (Figures 3 and 4). In the northern part of the Smoke Holes, Cave Mountain divides into several parallel ridges, each showing a maximum topographic relief of nearly 750 feet, which gives a unique geometric outline to the Smoke Holes: the region tapers along strike to the southwest (Figure 3).

REGIONAL STRUCTURE

The Smoke Hole region shows northeast-trending folds and faults. The main, major fold is Cave Mountain anticline (Figures 3, 4 and 5). Both the southeastern and northwestern limbs of the Cave Mountain anticline are broken by several reverse faults. Most minor folds are results of drag by the reverse faulting. Southeast-dipping reverse faults predominate, with an occasional, local, northwest-dipping reverse fault. Tilton (1927) mapped one of the more prominent reverse faults near Big Bend. Dennison (1955), Dennison and Naegle (1963), Tilton, et al (1927), and Perry (1971) have recognized similar faulting surrounding this region of the Valley and Ridge Province.

At present no deep wells have been drilled along the Cave Mountain anticline. To the west, along the Wills Mountain anticline, several deep wells have penetrated major thrust faults. From the data recorded by these deep wells, Rodgers (1963) and Gwinn (1964) agree that the main fold of the region (Wills Mountain anticline) is developed above a decollement. The fold developed where this decollement rises from the Cambrian Rome Formation to the Ordovician Reedsville Formation (Perry, 1971). The Cave Mountain anticline is associated with, if not the result of, a major, splay thrust fault developed off this rising decollement.

The reverse faults observed within the Smoke Holes are considered high-angle thrust faults. These faults are believed to be splays developed off the major thrust fault associated with the Cave Mountain anticline. Therefore, these faults are considered splay thrusts representing listric surfaces associated with the previously mentioned subsurface decollement. The faulting throughout the Smoke Holes will be referred to as thrust faulting. The northwest-dipping faults are considered local back-thrust development in relation to the southeast-dipping thrusts.

The Cave Mountain anticline is a complex anticline broken by thrust faults and trending N34^{o}E (Figure 3). The width of the fold is approximately two miles, as measured from the Oriskany Sandstone of Devonian age along the southeastern flank to the unbroken Oriskany Sandstone of Devonian age along the northwestern flank. The maximum structural relief of the Cave Mountain anticline is nearly 2775 feet. Some of the strata associated with tighter subsidiary folds are slightly overturned to the northwest.

The anticline is breached in the middle part of the Smoke Holes near Big Bend, where the Tuscarora Sandstone is the oldest formation exposed within the core of the fold (Figures 4 and 6). The fold plunges more steeply to the northeast (20-25^{o}NE) than to the southwest (8-15^{o} SW) of Big Bend. The trend of the Cave Mountain anticline varies, which was controlled by varied northwestward movements of thrusted blocks separated by cross-faults. The Cave Mountain anticline is narrower at the southwest end of the Smoke Holes than at the northeast end,

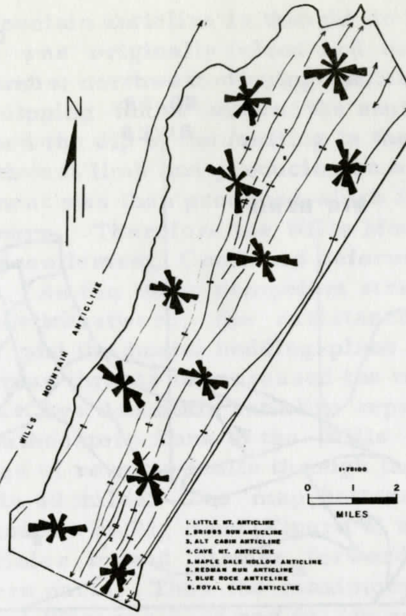

Figure 5. Map showing traces of axial surfaces along with joint rossette patterns. Joint analysis conforms to regional Appalachian trends. More prominent folds are labeled.

where the anticline divides into several folds that plunge northeastward into the Bedford syncline.

In the southern part of the Smoke Holes, the Cave Mountain anticline shows tighter major folding than it does near Big Bend in the center part of the Smoke Holes. A culmination is visible at Big Bend, where thrust faults are less common than at either end of the Smoke Holes (Figures 3 and 4). The Cave Mountain anticline, in the southern part of the Smoke Holes (A-A'), shows 54 percent compression of the Upper Ordovician through Lower Devonian section. In the northern part of the Smoke Holes (C-C'), a 45 percent compression of the section occurred while in the middle part (B-B'), a 64 percent compression of the section occurred. Thus an average compression of 54 percent of the section occurred throughout the Smoke Holes, as shown by the cross-sections (Figure 4).

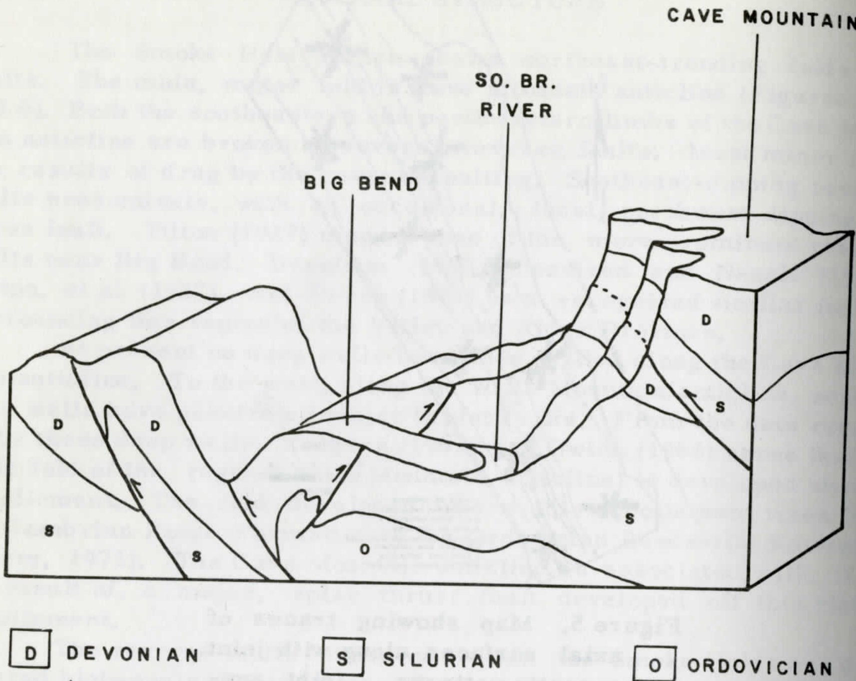

Figure 6. Geologic Block diagram of central part of the Smoke Hole Region. Southeast dipping high-angle thrust faulting is observed along with associated northwest-dipping back-thrust.

FAULTS

Detailed geologic mapping has shown high-angle thrust faulting with some back-thrust development and associated cross-faulting along the Cave Mountain anticline. Drag folding is prevelant along the thrust faults. The maximum stratigraphic throw along the thrust faults is nearly 1900 feet at Big Bend where the Mifflintown Formation of Middle Silurian age has been faulted over the Oriskany Sandstone of Lower Devonian age (Figures 4 and 6). The fault surfaces, measured at ground level, generally dip 50-55° SE and are interpreted to represent the diagonally-upward shearing of splay thrusts showing listric surfaces. These longitudinal faults (thrust faults) are considered to become bedding plane thrusts within the subsurface. In the Smoke Holes, thrusting is concentrated within the Middle Devonian formations and the Upper Silurian limestones and shales. Most of the structure shows the lower Helderberg Group thrust over the upper Helderberg Group and the Oriskany Sandstone. With increased depth of erosion, the Tonoloway Limestone of Silurian age is shown thrust over the Helderberg Group.

The Wills Mountain anticline is thought to have had a structural history in which it was originally folded and then broken by reverse faults within both limbs; northwest-dipping faults within the northwest limb and southeast-dipping faults within the southeast limb. Further deformation steepened the dip of the faulting in the northwest limb while overturning the northwest limb and producing an asymmetrical fold. A subsurface decollement was then produced which folded the Wills Mountain anticline still more. Therefore the Wills Mountain anticline is the major fold above a decollement. Continued deformation produced splays off this decollement. As the more competent strata above the decollement were pushed northwestward, the resistance to forward motion gradually increased and the initial bedding-plane thrusts broke upward producing several splay thrusts that crossed the more competent strata at high angles. The Cave Mountain anticline represents this "piling of strata" along the southeastern flank of the Wills Mountain anticline. Individual map traces of reverse faults through the Smoke Holes extend along strike for 20 to 40 miles. The map pattern of these longitudinal faults and the culmination at Big Bend (Figure 4) suggest that the middle part of the Smoke Holes moved further forward and upward than the northern and southern parts. Thus the maximum faulted displacement is found at Big Bend (Figures 3, 4 and 6). In general, the structural pattern of the area is one of multiple, subparallel to imbricate faulting with associated drag folding.

At Big Bend there are major conjugate high-angle thrust faults, one dominant over the other (Figures 4 and 6). The northwest-and the southeast-dipping thrust faults both dip about 54°. The northwest-dipping back-thrust extends nearly 1.5 miles along strike and shows a maximum stratigraphic throw of approximately 100 feet. This small displacement is to be expected, because horizontally compressed areas generally fail along shear fractures dipping toward the source of compression rather than along those dipping away from the source of compression. Less pronounced back-thrusting occurs in the southeastern limbs of several of the smaller folds.

Throughout the Smoke Holes several fault zones are visible. These fault zones were noted by the attitude of bedding, stratigraphy, springs, eroded areas, crushed and tightly folded beds, mineralization, vegetation, and topography. Folds and subsidiary faults often accompany a major fault, and the resulting shattered zones are generally more susceptible to weathering and erosion. Breccias, containing many angular fragments of the beds involved in the faulting, were found in zones as much as 1.5-2.0 feet thick. Faulting within the sandstones produced shattered zones and metaquartzites. Psilomelane was found along the fault zones, indicating post-fault mineral deposition. The limestones and shales are characterized along the fault zones by highly complex folding and well-developed axial-plane cleavage.

Associated with the Cave Mountain anticline are several northwest-trending, strike-slip, cross-faults (Figure 3). As the cross-faults seem to offset the thrust faults, the cross-faulting must have occurred during or after thrusting. Some cut perpendicularly across the fold axes and thrust surfaces while others cut diagonally (at 10-15°) across the fold axes. The cross-faults show horizontal, usually left-lateral net-slip with occasional vertical displacement. The maximum net-slip along the cross-faults is approximately 100 feet. Few cross-faults occur in the central Smoke Holes, at the culmination of the Cave Mountain anticline at Big Bend (Figure 3). A concentration of diagonal right-lateral faults occurs in the southern area between major left-lateral cross-faults. This cross-fault zone may be the result of the upward growth of the thrust block under High Knob. From continued lateral compression, left-lateral tear faults developed by further forward motion of the thrust block to the north as compared to the area south of Cave Mountain Gap. This cross-fault concentration zone suggests the possibility of a lineament through this region. This zone is most apparent on aerial photographs and is reflected on the ground-surface by several fracture traces, water gaps, drainage patterns, plunging structures and reduced topography. There is noticeable similarity found at the northern end of the Smoke Holes where the Cave Mountain anticline plunges into the Bedford syncline.

FOLDS AND JOINTS

Most major and minor folds trend about N32°E (Figure 5). Many are interpreted as drag-folds caused by movement on the thrust faults. Most such folds are asymmetrical, overturned to the northwest, with axial surfaces dipping from 55° SE to vertical. Many of the axial surfaces are irregular in dip due to slip on bedding planes or fractures that occur in the various formations involved in the folding. The amplitude of some of these folds is as much as 2500 feet. Cheveron folds are developed within some of the more closely spaced thrust belts. Several of the more prominent folds throughout the Smoke Holes are: Little Mountain anticline, Briggs Run anticline, Alt Cabin anticline, Maple Dale Hollow anticline, Ship Rock syncline, Blue Rock syncline, Redman Run anticline, Blue Rock anticline and Royal Glen anticline (Figure 5). These anticlines are accompanied by smaller, usually unnamed, synclinal drag-folds. Because of the greater amount of plunging, closeness of folding is not exposed in the northern area. Whereas in the southern area, tighter folds are shown within an equivalent stratigraphic sequence to that of the northern area. By superimposing the map traces of the fold axial-surfaces (Figure 5) on the geologic map (Figure 3) the relationship of the folding to the faulting may be noted: the folds are associated with the faults. The anticlinal drag-folds occur on the upthrown block of the faults.

The central parts of both Cave Mountain and Wills Mountain anticlines are structurally elevated, at the Big Bend culmination, with

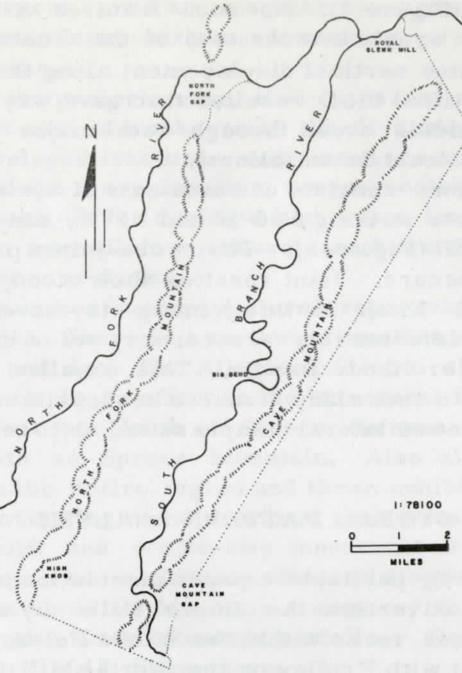

Figure 7. Drainage and prominent topographic features of the Smoke Hole Region, West Virginia. Notice the strong eastward concavity of the topography and the drainage, along with the central meander belt of the South Branch River.

respect to the northern and southern ends. Such observations suggest that the area of the Big Bend culmination moved northwest more readily than did the southern and northern ends of the Smoke Holes. The greater movement of the middle part of the Smoke Holes brought about the increased upward growth of that part of the structure, thus producing maximum displacement and release of stress along the northwest-dipping back-thrust at Big Bend.

The middle part of the Smoke Holes shows curving map trends of axial-surfaces, convex eastward (Figures 3, 5 and 7). This curve is reflected in topography, geology and stream patterns of the North Fork and South Branch Rivers of the Cave Mountain and Wills Mountain anticlines. The lowest topography along both North Fork Mountain and Cave Mountain is located in this middle part and within this part the stream patterns are convex eastward and show well-developed meanders as compared to their normal, youthful, straight segments in the northern and southern parts (Figure 7). No such features are observed within the Middle Mountain syncline to the east of the Smoke Holes. It is inferred that, with some vertical displacement along the bounding cross-faults, a southeast tilted block resulted that gave way to those features observed for the middle areas through both major structures (Wills Mountain and Cave Mountain anticlines).

A joint system consists of three sets of systematic joints: two sets of diagonal joints striking N60°W and N5°W, and one set of cross-joints striking N31°E (Figure 5). The cross-joints predominate where more open folding occurs. Joint rosettes show a conjugate angle of approximately 55-60°. Much of this jointing may have developed during the early stages of deformation. A secondary set of diagonal joints appears, with a smaller dihedral angle. This smaller set of fractures may have been due to secondary stresses responsible for the associated cross-faulting, renewed lateral compression, or to regional relaxation.

STREAM PATTERN ANALYSIS

Fridley (1939), published a paper on solution and stream piracy of the South Branch River into the Smoke Holes by underground solution of the calcareous rocks within the Smoke Holes. The author is in complete agreement with Fridley on the North Mill Creek valley (that valley between Cave Mountain anticline and Middle Mountain syncline) being the abandoned channel of the South Branch River, which has definately been diverted into the rough, mountainous Smoke Hole Region from a broad, mature stream valley.

Fridley's piracy by underground solutional chambers is a possibility, but leaves question as to the actual cause of diversion. To have solutional piracy, the river would have to have eroded down to, or very near to, the Helderberg Group in order to have the underground chambers intersect the surface, thus allowing an escape route for the river. From the lowest abandoned river deposits, the stream was at a higher level than at the present time. The river's present course through the Little Mountain Gap, capped by Oriskany Sandstone, is presently exposing the top of the calcareous Helderberg Group. The river's old channel is in the Devonian Shales and diversion by solution would require a solutional channel through the Oriskany Sandstone and into the underlying limestones. Because the ancient river grade was higher, then at that time the Oriskany Sandstone had not been eroded down through to the limestones. Therefore, another means was probably responsible for the initial piracy into the Smoke Holes. Granted, once within the Smoke Holes, solutional chamber flow could have been possible, but was probably limited to channels along the structural strike and not across the strike as it would have had to have been to develop the initial piracy channel by solution. I have visited a majority of the caves in this area that represent the former solutional channels. The majority and the largest caves are singular passages, developed along the strike of the structures, with no branching or network channels perpendicular to strike.

By observing the geologic map, it is apparent that there is a northwest-southeast zone of weakness at the diversionary area, which consists of well developed fractures and weak cross-faulting zones. Erosion along one of the more dominant zones created a trough to capture the South Branch River during a high water level and the stream was diverted into an immediate northwest-southeast flow influenced by these fractures and cross-faults.

Actually the cross-structure features represent a major lineament through the area. There is a northwest-southeast drainage through the entire region and into the Allegheny Structural Front. Directly along this zone there is an offset in the Structural Front and a water gap through the Front at Mouth of Seneca (Seneca Rocks); the Front continues to the south as Spruce Mountain. Also along this zone are the highest areas in the entire region and these exhibit a pseudo-radial drainage pattern, indicating some dominant cross-faulting feature under the surface with uplift and strike-slip movement or some subsurface structural high. Two other noticeable features directly along this zone are: (1) an offset in the Wills Mountain structure, and (2) a narrowing or tapering of the entire Cave Mountain anticline. Perhaps this represents one of Gwinn's (1964), lineaments.

Field observations and a Chi Square test suggest that the South Branch River has been diverted from the mature North Mill Creek stream valley to a youthful stream valley by systematic joints and cross-fault zones (Sites, 1971). Those influencing factors compared with the stream pattern segments were: (1) thrust fault traces, and (2) joint and fracture traces. A Chi Square test was performed on each set and the results showed that the segments of each set did have a preferred orientation and were not randomly developed. An analysis for a vector mean of each set was computed to establish the preferred orientations. The results showed that once in the Smoke Holes, the stream's pattern is structurally controlled by the longitudinal faulting. The river then flows northeastward along strike, with minor deviations attributed to joints and cross-faults. In the middle part of the Smoke Holes the stream appears to be controlled by the Big Bend culmination, cross-faults, dip of strata and lithology. North of the culmination the stream's pattern again is influenced by the longitudinal fault zones.

SUMMARY

The doubly-plunging Cave Mountain anticline shows numerous southeast-dipping, high-angle thrust faults and associated cross-faults whose movement was dominantly strike-slip. Major, highly asymmetrical to overturned folds occur throughout the Smoke Holes. Depending upon resistance to forward motion, some parts of the structure moved further northwestward than others, resulting in curving map trends of axial surfaces. A northwest-dipping back-thrust is locally developed within the middle part of the Cave Mountain anticline where maximum forward displacement (thrust) occurs along with a structural culmination. More tightly folded cores of anticlines are exposed within the southern part of the region due to steeper plunging of hinge lines within the northern part. The southeast-dipping faults are considered third-order splay thrust faults that, in the subsurface, merge into a thrust within the Ordovician shales. This thrust represents a second-order splay from the underlying decollement in the Cambrian Rome Shale. Preferred orientations of systematic joints are relatively consistant throughout the area and conform to regional Appalachian trends. The pattern of the South Branch River appears to be structurally controlled by the longitudinal faulting through the Smoke Holes.

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STRUCTURAL SIGNIFICANCE OF KYANITE-SILLIMANITE QUARTZITES IN ROWAN AND IREDELL COUNTIES, NORTH CAROLINA

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ABSTRACT

Kyanite-sillimanite quartzites, probably representing a northeastern extension of the Kings Mountain belt, are exposed in the Central Piedmont in Rowan and Iredell Counties, North Carolina. Prior to this study, the Kings Mountain belt was believed to end eighteen miles to the west in Iredell County.

The quartzites crop out discontinously for at least twenty and possibly twenty-six miles. The most prominent exposure is eight miles NW of Salisbury on a ridge 0.9 mile long and 0.2-0.3 mile wide; the beds striking N 35°-65°W and dipping to the southwest. Isolated quartzites crop out about 1 and 10 miles from the northwestern end and 2-4 miles and 10-11 miles from the southeastern end. A connecting link occurs in eastern Iredell County, ten miles west of the ridge and eight miles northeast of the northern termination of the Kings Mountain belt as presently mapped. The bedding here strikes N 35°E and dips 80°NW, and abundant quartzite float and large boulders of sillimanite-muscovite quartzite are found in an area several hundred feet long.

The quartzites consist of (1) granular orthoquartzites and muscovite quartzites, (2) porphyroblastic kyanite quartzites, (3) lepidoblastic kyanite-sillimanite quartzites, and (4) minor chevron-folded muscovite schists. Accessory minerals are rutile, pyrite, and rare staurolite. Needles of rutile inside the quartz and small tetragonal rutile crystals characterize the unit.

The quartzites exposed in Iredell and Rowan Counties are probably eastern-northeast extensions of the Kings Mountain belt. The Kings Mountain belt was once much more extensive, assimilation by Charlotte belt intrusions and later erosion have reduced the belt to isolated outcrops of resistant quartzites.

INTRODUCTION

Recently discovered and previously unreported orthoquartzites, mica quartzites, kyanite and kyanite-sillimanite quartzites and minor mica schist crop out in the Piedmont of western Rowan and southeastern Iredell Counties, North Carolina. The quartzites are probably extensions of the Kings Mountain belt previously considered to terminate in Iredell County at Duke Power State Park (Conrad, 1964), eighteen miles to the west.

The purpose of this report is to describe the occurrence, distribution, mineralogy and petrography of the quartzites, and to discuss their origin.

Acknowledgments

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GENERAL GEOLOGY

The geology of the Central Piedmont of North Carolina is not completely understood. On the basis of differences in metamorphic rank, lithology and percentage of igneous intrusive the Piedmont has been divided into a series of northeast trending belts. In Central North Carolina these are from west to east: Inner Piedmont, Kings Mountain belt, Charlotte belt, and Carolina slate belt (King, 1955). The Kings Mountain belt is a metamorphic complex of amphibolites, marbles, schist, and quartzites cut by swarms of spodumene pegmatites. The Inner Piedmont to the west is represented by a sequence of kyanite-biotite gneisses and schists containing some concordant monzonite intrusives. The Charlotte belt is represented by various metamorphic rocks intruded by younger granites, quartz monzonites, gabbros, syenite, and diabase dikes. Butler and Ragland (1969) divided the intrusives of the Piedmont into premetamorphic and syn- or post-metamorphic. The quartzites described herein crop out in both the western Charlotte belt and the eastern Inner Piedmont belt.

OCCURRENCE

The quartzites crop out discontinously for at least twenty and possibly twenty-six miles (Figure 1). The most prominent exposure is on a ridge 0.9 mile long and 0.2-0.3 mile wide, here the beds strike N35°-65°W and dip to the southwest. The outcrop is located eight miles

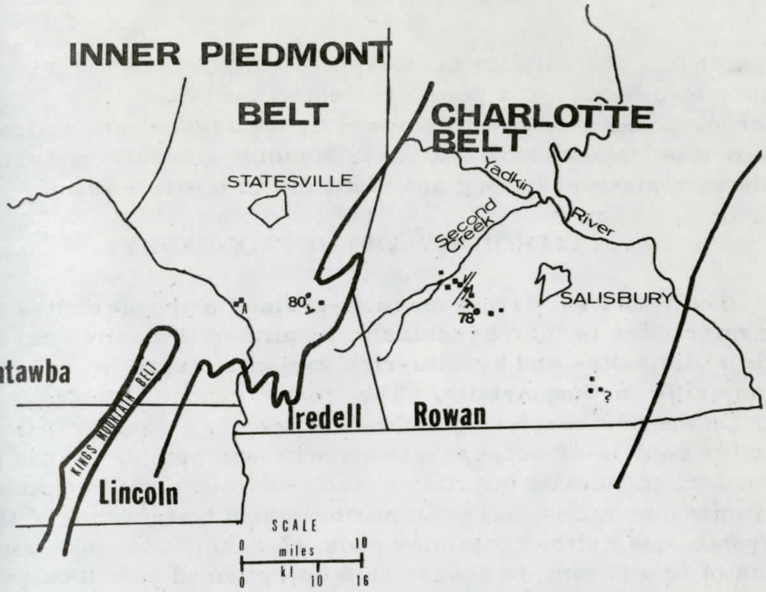

Figure 1. Outline map showing the locations of the quartzite outcrops (small squares); the A designates the location of the quartzites described by Conrad (1964). Base map and contacts from the Geologic Map of North Carolina.

NW of Salisbury at the end of county road 1733. The ridge rises about 100 feet above the surrounding terrain and ends abruptly 160 feet above Second Creek. Poorly exposed isolated quartzites crop out along the projected strike of the ridge about 1 and 10 miles from the northwestern termination and 2-4 miles and 10-11 miles from the southeastern end of the ridge. A connecting link is located in eastern Iredell County, ten miles west of the ridge and eight miles northeast of the northern termination of the Kings Mountain belt. Abundant quartzite float and large boulders of sillimanite-muscovite quartzite are found in an area several hundred feet long. The beds strike N 35° E and dip 80° NW. Petrographic and mineralogical similarity suggests that these quartzites are extensions of the same unit.

The ridge is arcuate in outcrop and trends N 30-35° W for about 0.3 mile then abruptly changes to N 65° W for the remaining 0.6 mile. The unit is faulted; a small fault striking N 70° E offsets the ridge crest about 30 feet, 1000 feet from the northwestern end. Talus and slump blocks cover most bedrock on the slopes, thus obscuring the attitude of the bedding. One outcrop near the northwestern termination that appears to be in situ contains interbedded coarse and fine-grained layers dipping 78° southwest.

The quartzite exposed along the ridge is weathered and covered by vegetation. The smaller boulders are soft, friable, and have the appearance and texture of a poorly cemented sandstone. The larger blocks are dense, coherent, and contain pyrite crystals, but are stained reddish brown in a selvage about 2 cm thick. Kyanite occurs in pods and stringers; these resist weathering and stand out in positive relief.

LITHOLOGY AND PETROGRAPHY

The rocks vary from medium-grained orthoquartzites and muscovite quartzites to porphyroblastic kyanite-sillimanite quartzites and include sillimanite- and kyanite-rich variants, rare mica schist and conglomeritic orthoquartzite. The rocks change lithology along the ridge. On the southeastern end they are mostly clean, granular, orthoquartzites containing accessory muscovite and pyrite. These grade into granular, muscovite quartzites and schistose sillimanite-kyanite and kyanite quartzites near the northwestern termination of the ridge. The kyanite quartzites containing pods (7-8 cm) and thin discontinous ribbons of fine (2 mm) to coarse (1.5 cm) grained pale blue to gray kyanite crystals. Tightly folded muscovite schists with 1.0 cm chevrons and flexural slip folds occur in mica quartzite float and are present in talus on the northwestern end of the ridge. Rare tourmaline crystal groups (1.5 cm wide) are found in float. The isolated outcrops to the northwest and southeast of the ridge are composed of orthoquartzites, muscovite-bearing orthoquartzites, and muscovite-sillimanite quartzites.

The quartzites vary widely in mineralogy and texture, but can be divided into five mineralogical types:

1. Orthoquartzite to muscovite-bearing quartzite 93-99 percent quartz with 1-7 percent muscovite and accessory pyrite and rutile.
2. Kyanite quartzite - 25-60 percent kyanites, 40-75 percent quartz, with accessory muscovite and rutile.
3. Sillimanite quartzite - 30-40 percent sillimanite, 60-70 percent quartz, with accessory muscovite and kyanite.
4. Sillimanite-kyanite quartzite - rock containing 20-30 percent kyanite and sillimanite plus quartz and accessory muscovite.
5. Muscovite schist - essentially muscovite with 10-15 percent quartz and pyrite. All of the pyrite is oxidized to limonite in the samples examined.

The orthoquartzites and kyanite-sillimanite quartzites are the most abundant. Textures vary from lepidoblastic in the sillimanite-rich rocks (Figure 2A) to granoblastic in the quartz-rich rock (Figure 2B). All constituents are fine to medium-grained except the porphyroblasts. The kyanite, sillimanite, and muscovite are set in a matrix of xenoblastic quartz (0.06-2.00 mm) that invariably shows undulatory extinction.

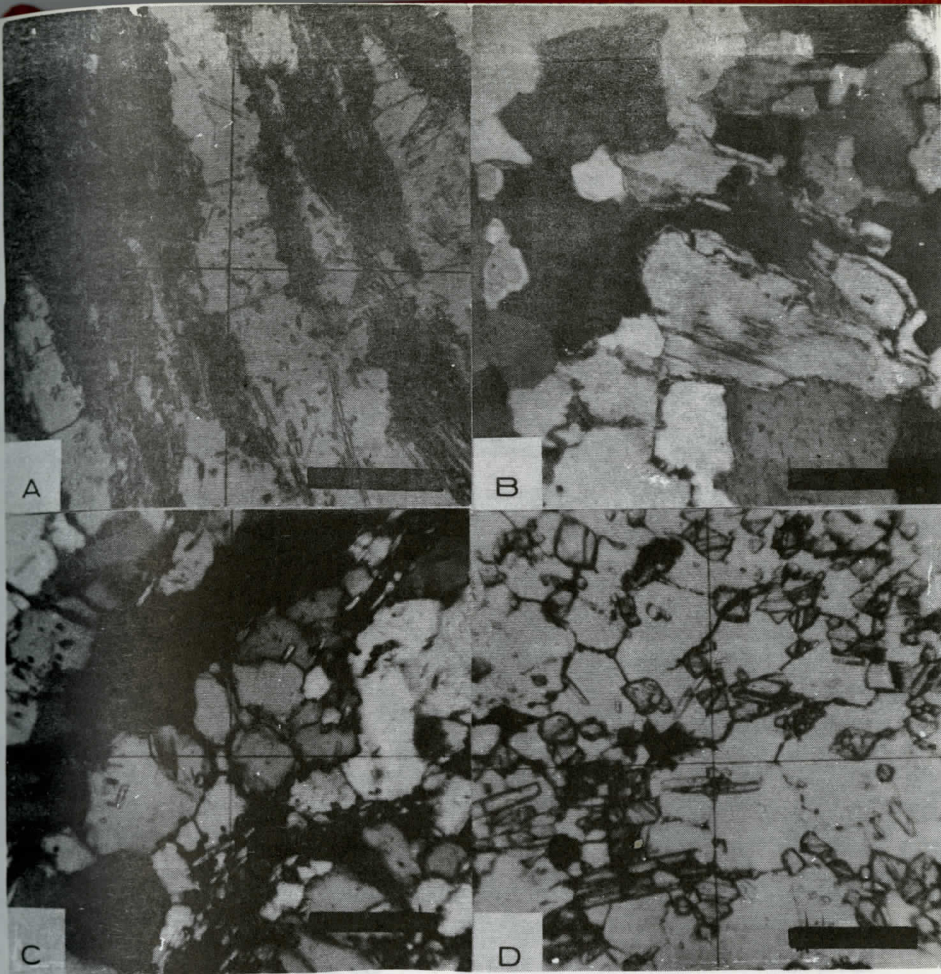

Figure 2. Mineralogy and texture of the quartzites, scale is 0.5 mm.
A. Lepidoblastic sillimanite quartzite. Plane light.
B. Granoblastic muscovite quartzite. Crossed nicols.
C. Sillimanite inclusions in quartz. Crossed nicols.
D. Sillimanite quartzite cut perpendicular to lineation. Plane light.

Sutured growth contacts are prominent between individual grains and much of the quartz contains inclusions of small sillimanite or rutile needles and tiny rutile crystals (Figure 2C). Muscovite (0.1-1.0 mm) is concentrated in thin separate lamellae 1-3 cm apart. Prismatic idioblastic sillimanite (0.3-0.6 mm) concentrated in thin (0.5-1.0 mm) layers produce a lepidoblastic texture. A preferred orientation of the sillimanite is obvious in sections cut perpendicular to the long axes of the sillimanite (Figure 2D).

Idioblastic to xenoblastic kyanite porphyroblasts (1-6 mm) in thin sections, up to 2 cm in hand specimens) occur as isolated crystals surrounded by quartz (Figure 3A) and as aggregates of larger crystals in kyanite quartzites (Figure 3B). The kyanite and sillimanite are

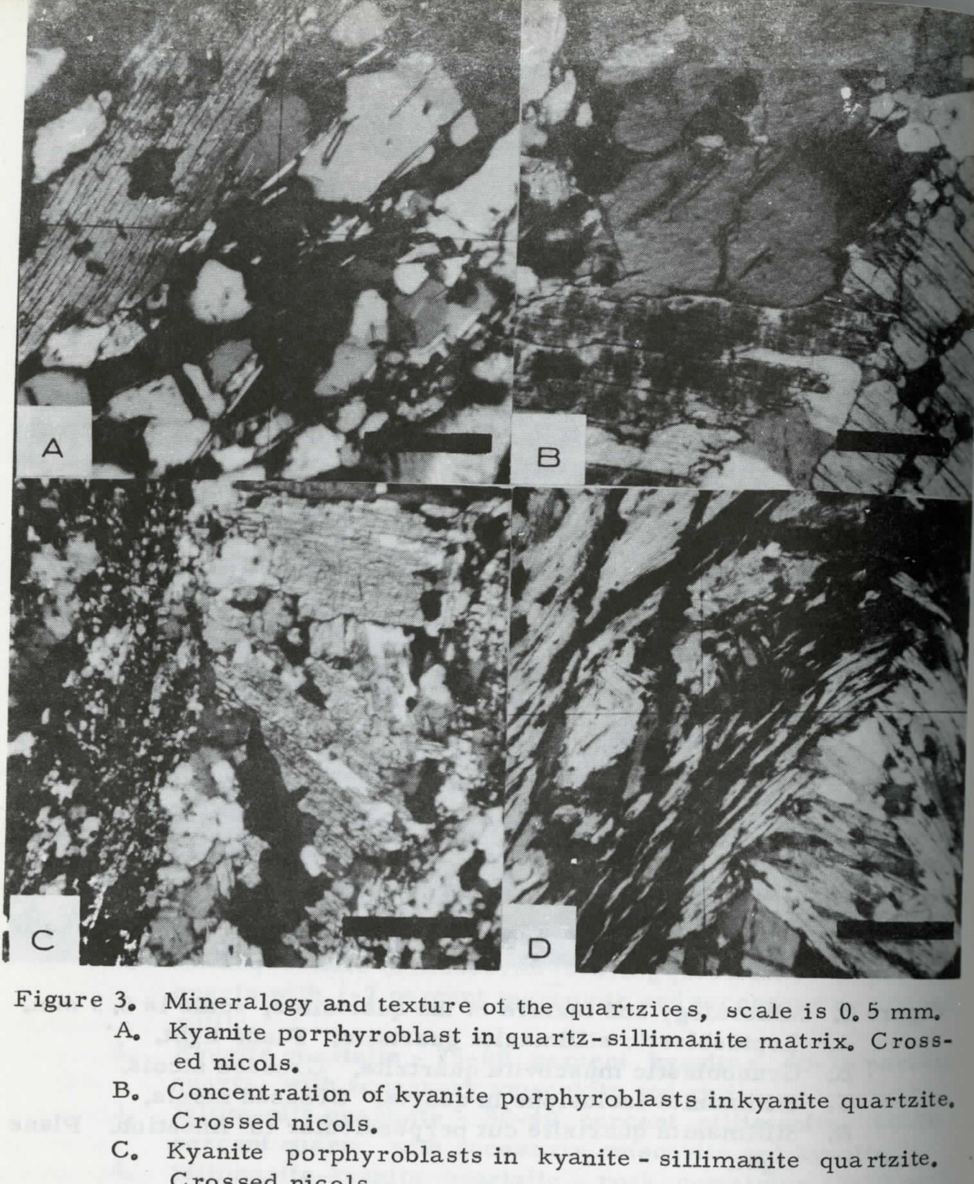

Figure 3. Mineralogy and texture of the quartzites, scale is 0.5 mm.
A. Kyanite porphyroblast in quartz-sillimanite matrix. Crossed nicols.
B. Concentration of kyanite porphyroblasts in kyanite quartzite. Crossed nicols.
C. Kyanite porphyroblasts in kyanite - sillimanite quartzite. Crossed nicols.
D. Chevron in muscovite schist. Crossed nicols.

commonly concentrated in alternating layers 1-3 mm apart (Figure 3C). Kyanite is slightly poikiloblastic, containing small quartz and muscovite inclusions. Accessory pyrite and rutile occur as randomly distributed grains. Rutile is present as distinct, idioblastic tetragonal crystals (0.07-0.3 mm) and as fine needles (0.02 mm) present inside the quartz. Rare poikiloblastic staurolite crystals are present.

Muscovite schist, found in float at the northwest end of the ridge, is composed of xenoblastic muscovite (0.5 mm) separated by thin discontinuous layers of quartz and limonite after pyrite. The rock has been folded into a series of tight chevrons (Figure 3D). Chevrons are also present in an orthoquartzite exposed on a small hill 0.7 mile southwest of the ridge; here thin (0.5 mm) layers outline complex, irregular chevrons (wave length 2-6 cm) 1-3 cm apart.

CONCLUSION

With available data it is difficult to accurately determine the history of these quartzites. The following hypothesis is presented: The quartzites exposed in Iredell and Rowan Counties are probably eastern-northeast extensions of the Kings Mountain belt, indicating that the Kings Mountain belt was originally much more extensive than now. The combined effects of assimilation by intrusives of the Charlotte belt and subsequent erosion have reduced the northeastern part of the belt to widely separated outcrops of resistant quartzites.

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